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Geology of the Caledonia Park Quadrangle Area, South Mountain, Pennsylvania

John L. Fauth

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by John L. Fauth

Cooperating Geologist
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CONTENTS

Page

Abstract 1

Introduction 2
  Purpose 2
  Location 3
  Procedure 3
    Field methods 3
    Rock names and colors 5
    Laboratory procedure 5
  Previous investigations 5
  Acknowledgements 8

Physiography 8
  General statement 8
  Landforms 11
  Drainage 11

Stratigraphy 12
  General statement 12
  Precambrian rocks 13
    Catoctin Formation 13
      Metabasalt 17
      Epidosite 18
      Greenstone 18
      Bluish metarhyolite 18
      Mottled metarhyolite 20
      Metarhyolite breccia 20
      Red porphyritic metarhyolite 21
      Phyllites 21
    Stratigraphic relations 21
    Thickness of the Catoctin Formation 24
  Cambrian rocks 24
    Loudoun Formation 27
    Weverton Quartzite 29
    Harpers Formation 33
    Antietam Quartzite 36
    Tomstown Dolomite 39
    Waynesboro Formation 41
    Elbrook Formation 42
    Conococheague Formation 44
  Ordovician rocks 47

Page
Stonehenge Formation 47
Rockdale Run Formation 49
Cretaceous (?), Tertiary-Quaternary deposits 50
Terrace gravels 50
Alluvium 51

Metamorphism 51

Structural geology 52
Introduction 52
Structural elements 24
Primary planar structures 54
Secondary planar structures 54
Flow cleavage (S_1) 54
Slip Cleavage (S_2') 57
Fracture cleavage (S_2) 58
Joints 59
Linear structures 61
Lineations of orientation and elongation 61
Intersection of planar structure 68
Folds 70
Folds on bedding 70
Folds on cleavage 76
Faults 77

Regional geology 80

Geologic history 83

Structural dynamics and mechanics 88

Summary and conclusions 93

References 96

Appendix A — Petrography 101
Petrography of the Catoctin Formation 101
Metabasalt 101
Epidosite 102
Bluish metarhyolite 104
Mottled metarhyolite 105
Metarhyolite breccia 106
Red porphyritic metarhyolite 106
Phyllites 107
Petrography of the Chilhowee Group 107

Page

Loudoun Formation 107
Weverton Quartzite 108
Harpers Formation 110
Antietam Quartzite 114

Appendix B — Geology of local areas 115
Southern part of Piney Mountain 115
Rocky Mountain area 116
Eagle Rock - Poke Hill area 116
Long Pine Run - Birch Run area 117
Conococheague Creek Valley 117
Piney Mountain west of Strasburgh School 118
Mountain Creek Valley 119

ILLUSTRATIONS

FIGURES

Page

FIGURE 1. Location of mapped area in relation to physiographic provinces of Pennsylvania 4
2. Classification of sandstones 6
3. Evolution of structural interpretations of South Mountain, Pennsylvania 9
4. Locations of some recent investigations in the Blue Ridge 10
5. Stratigraphy of the Blue Ridge from northeastern Tennessee to Pennsylvania 16
6. Composition of the Chilhowee Group in south-central Pennsylvania 25
7. Schematic diagram of tectonic coordinates and major structural elements 53
8. Equal-area projections of cleavage 55
9. Relation between composition and flow cleavage development in Chilhowee Group 56
10. Joint pole projections 60
11. Lineation diagrams 62
12. Development of flow cleavage 65
13. Relation of 'tailed" quartz grains to rock composition and structural attitude 68

Page
14. Fold axes and faults in the Caledonia Park area 71
15. Quartz fabric diagrams 73
16. Geologic history of the Caledonia Park area 84
17. Structural development of the Caledonia Park area 98
18. Geology of the Chinquapin Hill-Yellow Ridge area...... 118

PLATES

Page
Plate 1. Geologic map of parts of the Scotland, Caledonia Park, and Arendtsville quadrangles, Pennsylvania ....pocket
2. Lower Cambrian sandstones and conglomerate and bedding and cleavage in Cambrian and Ordovician rocks 123
3. Photomicrographs of grains elongated parallel to flow cleavage; and cleavage 125
4. Bedding folds and quartz grain deformation in Antietam Quartzite 127
5. Minor bedding and cleavage folds and photomicrographs of the Catoctin metabasalt 129
6. Photomicrographs of the Catoctin metarhyolite and Loudoun Formation 131
7. Photomicrographs of the Weverton Quartzite, Montalto Quartzite, Harpers Formation and Antietam Quartzite 133

TABLES

Page
Table 1. Generalized stratigraphic column in the Caledonia Park area 14
2. Summary of sedimentary characteristics in the Chilhowee Group 26
3. Point count data - Catoctin metabasalt 103
4. Chemical analyses of Catoctin metabasalt from South Mountain, Pennsylvania 103
5. Point count data - Loudoun Formation 108
6. Point count data - Weverton Quartzite 111
7. Point count data - Harpers Formation 113
8. Point count data - Antietam Quartzite 114

GEOLOGY OF THE CALEDONIA PARK QUADRANGLE AREA SOUTH MOUNTAIN, PENNSYLVANIA

by

John L. Fauth*

ABSTRACT

The detailed stratigraphy and structure of the rocks in and adjacent to South Mountain in the Caledonia Park area of south-central Pennsylvania were established by field study and mapping. The area lies mainly in the Blue Ridge physiographic and tectonic province. Major rock units include: Precambrian rhyolites and basalts, Lower Cambrian graywackes, sandstones, and quartzites, and Cambrian and Ordovician carbonates. The Precambrian volcanic sequence is composed of intercalated, sub-aerial rhyolites and basalts deposited on a Precambrian granitic terrane. Thick widespread sediments of a shelf facies accumulated in a shallow sea and transgressed the volcanic rocks.

Folding is the dominant mode of deformation. South Mountain is the western limb of the Catoctin Mountain-Blue Ridge anticlinorium in the Caledonia Park area, and consists of a series of northeast-trending concentric and similar folds asymmetric to the northwest. Flexure and cleavage (shear) folds were developed depending upon rock competency, with shear folding effected along axial plane flow cleavage. The shear folding affected nearly all rocks during the last stages of deformation. Petrographic evidence and the existence of downdip lineation on the flow cleavage indicate that the principal movement in cleavage folding was the combined product of elongation parallel, and shortening normal to, the cleavage surfaces, coupled with slip between microlithons. A steep-dipping slip cleavage developed subsequently in conjunction with folding of flow cleavage in the phyllites. The primary stress field during these phases of deformation has P_1 (principal stress direction) oriented in a northwest-southeast direction.

Several faults were mapped including a right-lateral transverse fault near Caledonia Park which offsets structural axes and separates blocks with different intensities of folding. This fault probably developed early in the folding phase and is related to a change in the trend of the Blue Ridge near Caledonia Park.

Longitudinal faults cut through developing anticlines as a result of the stretching of the overturned limbs or the outward pressure within the narrowing core of the developing fold.

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Several structures formed in response to secondary stress fields. Local subhorizontal axial-plane cleavage and the associated small, recumbent folds in some beds of the Elbrook Formation are interpreted as gravity-collapse structures. These formed in a reoriented stress field in which P_1 was vertical.

Longitudinal extension resulted from a stress field with P_3 northeast-southwest and produced the plunge and arcuate plan of the South Mountain fold in Pennsylvania. Associated with a continued northwest-southeast oriented P_1, numerous transverse faults of steep dip developed in the latter stages of deformation.

Basement involvement in the deformation and the lack of important decollement zones are evident by the uniform deformation plan which affects all rocks in the stratigraphic section north of the Potomac River, including the Precambrian gneisses.

South Mountain structures, identified in rocks no younger than lower Middle Ordovician, are truncated by Triassic faults. Deformation is Paleozoic and may be related to the Taconic, Acadian, or Appalachian diastrophism. Tentative correlation of flow and slip cleavage in the study area with similar structures associated with dated rocks within the Piedmont Province, and with dated thrusting nearby in the Valley and Ridge province lends some support to a Late Devonian (Acadian) orogenic period. The rocks of the Caledonia Park area underwent regional metamorphism during Paleozoic deformation and are in the greenschist facies of metamorphism.

INTRODUCTION

PURPOSE

The geology of the South Mountain area of Pennsylvania was last studied during the early part of the century as published by Stose (1909, 1929, 1953). Except for studies related to the stratigraphy of the Cambo-Ordovician carbonate sequence, little geologic research has been done in this area since that time. However, within the past decade, detailed geologic work in adjacent areas of Pennsylvania, Maryland, and Virginia has resulted in the recognition of significant structural complexities not evident in earlier reports.

Discoveries such as large-scale recumbent folding and associated faulting in Pennsylvania (Gray, and others, 1958) and complex cleavage folding in the Blue Ridge of Maryland and Virginia (Nickelsen, 1956) stimulated interest in the geology of the Pennsylvania segment of the Blue Ridge and its relation to the tectonic style of nearby areas. In addition, a study by the writer (Fauth, 1962) on a limited portion of South Mountain suggested the need for modification of existing local structural and stratigraphic interpretations.

More specifically, the objectives of the present study include:

(1) the investigation of the petrographic character and stratigraphy of the Precambrian Catoctin volcanics and the overlying Chilhowee group,

(2) a comparative study of the structure elements developed in an area of diverse rock type,

(3) the detailed mapping and the interpretation of the bedrock geology of the area,

(4) the evaluation of the Blue Ridge-Cumberland Valley physiographic boundary in terms of the differences in structural style on either side,

(5) the interpretation of these data in terms of structural genesis and the tectonic evolution of the area against a background of regional tectonics.

LOCATION

The area of the report lies in southcentral Pennsylvania, approximately midway between Chambersburg and Gettysburg (Figure 1). It includes portions of Adams, Cumberland, and Franklin counties. The map area is about 96 square miles, of which 16 square miles were originally studied in conjunction with the writer's previous report (Fauth, 1962).

The area is bordered on the north by 40°00′N., on the west by 77°35′00″W., and on the south by 39°52′30″N. On the east, the boundary is the line trending N46E and passing through Wenksville near the north end and the junction of U.S. 30 and Pa. 234 toward the south. The area lies athwart portions of two physiographic and structural provinces (Figure 1). The Blue Ridge province extends from northern Georgia to southeastern New York and is paralleled on the west by the Valley and Ridge province. South Mountain, the name given to the Blue Ridge in northern Maryland and Pennsylvania, is bounded on the west by the Cumberland Valley, the local segment of the Great Valley section, and on the east by the Triassic Lowland section. South Mountain extends northeastward to the vicinity of Carlisle, about twenty miles from the area of this report.

PROCEDURE

Field Methods

About 48 weeks were spent in the field during the summers of 1960-1962 and the spring of 1963. Field data were plotted directly on U. S. Geological Survey topographic maps of the Arendtsville, Caledonia Park, and Scotland 7½′ quadrangles (scale 1:24,000). Topography, culture, altimeter readings, and air photos were used for locating field positions. Air photos were also employed to locate outcrops, discover structural patterns, and trace topographically expressed beds across areas of poor exposure.

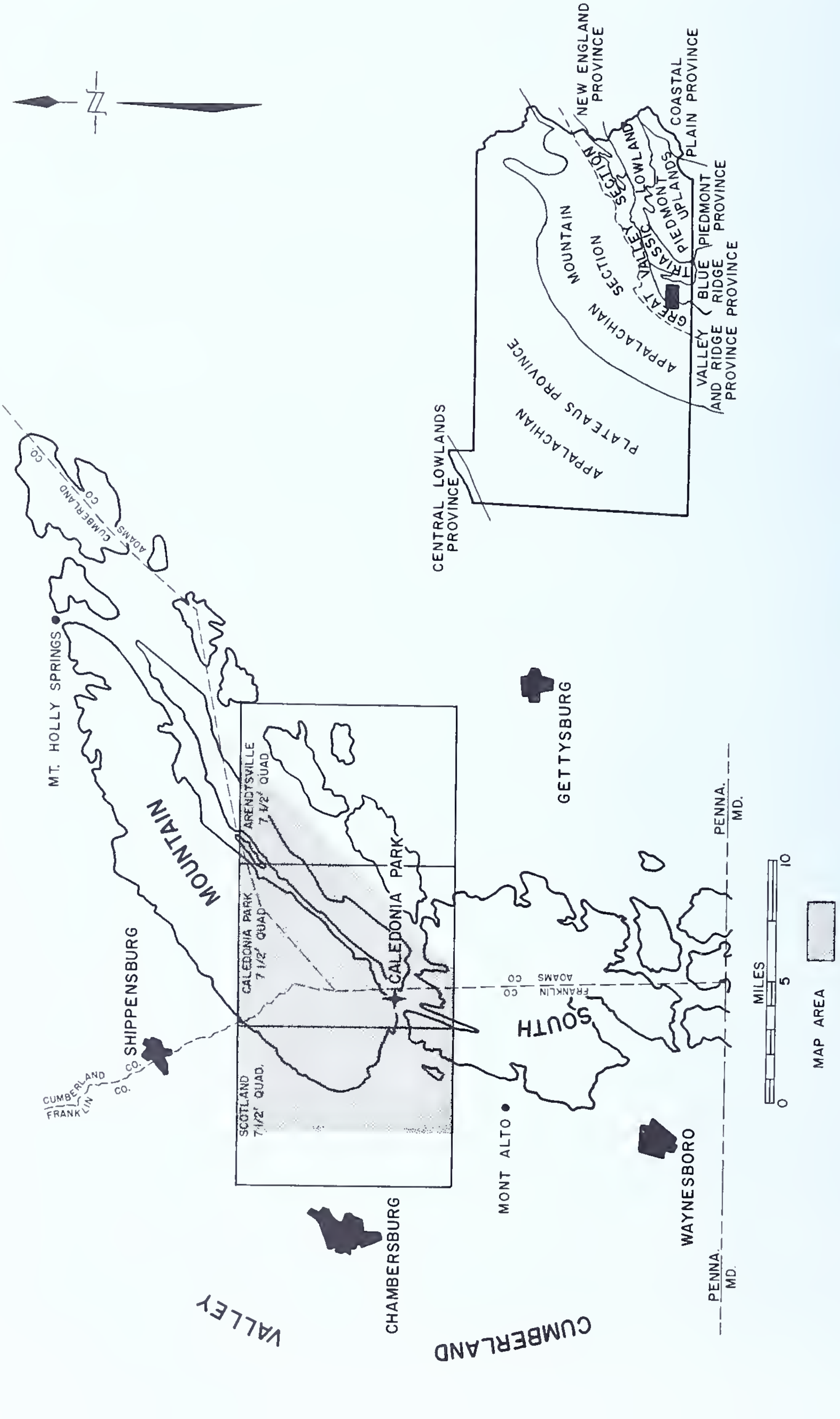


Figure 1. Location of mapped area in relation to physiographic provinces of Pennsylvania.

Rock Names and Colors

Because the rocks of the area have undergone only low-grade metamorphism, sedimentary and igneous rock names are used. The sedimentary rock names are based on the classification of Krynine (1948), illustrated in Figure 2. The term "phyllitic", however, has been used to further describe those sandstones which have a visible foliation. This foliation, found in rocks which otherwise retain most of their megascopic sedimentary characteristics, is commonly recognized by a silky sheen imparted to the surface of the cleavage by micaceous minerals. Metamorphic alteration of the volcanics is acknowledged by the use of the names metabasalt and metarhyolite.

Symbols following rock colors refer to specific colors in the Rock Color Chart (Goddard, and others, 1948); where such symbols are absent, color names do not necessarily comply with this chart. Grain size was determined by visual comparison with sieved standards.

Laboratory Procedure

Measurements of planar and linear structures taken in the field, and lineations (i.e., intersection of bedding and cleavage) as determined by stereographic methods (Nevin, 1949, p. 382-384) were plotted on the southern half of equal area nets.

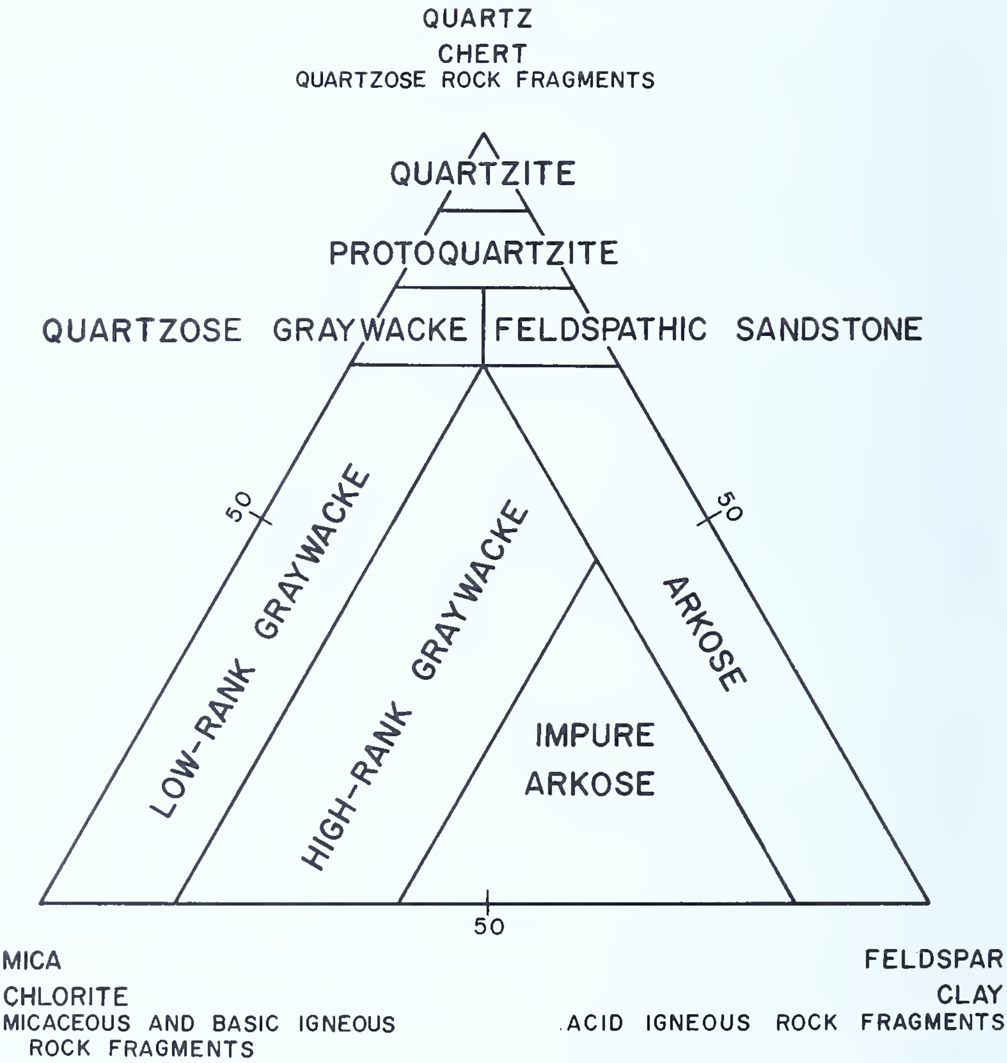
A suite of about 175 thin sections representing most of the exposed formations was examined. Etched section of carbonates and heavy mineral suites were also studied.

Unless otherwise noted percentage mineral compositions are visual estimates based on charts prepared by Terry and Chilingar (1955, p. 229-234). Point counts include 500 points per slide.

PREVIOUS INVESTIGATIONS

The initial investigation and report on the geology of the South Mountain region of Pennsylvania was a product of the work of H. D. Rogers (1858). Bascom (1896, p. 16) noted that "many of the characteristic features of the South Mountain rocks were aptly described by Professor Rogers, yet, plainly, their nature was not fully understood nor their importance appreciated . . ."

P. Frazer (1876; 1877) re-examined South Mountain, studied the stratigraphic sequence and interpreted its broader structural features. Frazer distinguished two classes of rocks: an older sequence of quartzites and arenaceous schists, and an overlying series of schists and orthofelsites,



MODIFIED FROM KRYNINE (1948)

Figure 2. Classification of sandstones.

both porphyritic and non-porphyritic. Neither of the orthofelsite types was recognized as an igneous rock by Frazer, and his structure cross-sections (Frazer, 1877, p. 285-299, sections 9-11) show an east-dipping homoclinal structure modified by local concentric folds. These stratigraphic and structural interpretations were held as late as 1892 by Lesley (1892, p. 144) who refers in his report to "several thousand feet of lower quartzite . . . and several thousand feet of an overlying feldspathic, micaceous, and chlorite series."

Both G. H. Williams (1892, p. 482-496) and F. Bascom (1893), working in the northern and southern extremities of South Mountain respectively, recognized two of the rock types as being volcanic. Bascom (1893; 1896) examined both the acidic and basic types, but concentrated on the petrographic study of the metamorphosed acid volcanic rocks, the aporhyolites.

In 1892 C. D. Walcott (1896, p. 24) discovered the trilobite *Olenellus* associated with the molluscan? *Hyolithes communes* in the quartzite series near Mount Holly Springs. Examination of the exposed section east of Waynesboro by Walcott and Keith (Walcott, 1896, p. 24-25) yielded fragments of *Olenellus* and specimens of the brachipod *Camerella minor.* The stratigraphic relationships of the rocks were established — the "orthofelsites" were older, and overlain by a quartzite series which was assigned an Early Cambrian age.

Keith (1893; 1894) mapped the Blue Ridge from northern Virginia to Pennsylvania, described the rocks, and divided them into formations. The structure was interpreted as anticlinal with the Lower Cambrian exposed in synclines in South and Catoctin Mountains. Stose and Stose (1946) corroborated Keith's interpretation of South Mountain structure in Frederick County, Maryland.

Stose (1906; 1932; 1953) and Stose and Bascom (1929) added information on the rock units described by Keith, and traced them northeastward into Pennsylvania. The South Mountain of Pennsylvania is described as "a great anticlinorium consisting of three or four anticlines and numerous thrust faults, bounded on the east by a series of normal faults along the margin of the Triassic rocks" (Stose, 1932, p. 80).

Cloos (1947; 1951) advanced the structural hypothesis that the Blue Ridge belt is an overturned anticlinorium with its overturned limb in South Mountain and its normal limb in Catoctin Mountain. The internal structure of South Mountain in Maryland was described as anticlinal. Cloos (1947) also attempted a quantitative measurement of the South Mountain deformation by measuring the distortion of oolites. Distortion as high as 100-150 per cent was found to be common.

Cloos' structural interpretation was confirmed by the work of Nickelsen (1956) and Whitaker (1955). Nickelsen (1956) mapped folding in South Mountain of a style and degree not previously recognized.

Since the pioneering investigations of Bascom (1893; 1896; 1897), work on the volcanic rocks associated with South Mountain has been centered in Virginia. The volcanic rocks are thought to represent an accumulation of plateau basalts by Reed (1955), whereas Bloomer (1950) suggested that the original rock was an andesite. However, both writers are in accord regarding the lack of a regional unconformity between the volcanics and the overlying meta-sediments.

Recent stratigraphic studies by Wilson (1952) and Sando (1957; 1958) in the Great Valley sections have resulted in revision and redefinition of units in the Cambro-Ordovician carbonate sequence and their extension into southcentral Pennsylvania.

A study by this writer (Fauth, 1962) noted differences in the stratigraphic framework and structural style in the Pennsylvania segment of South Mountain compared to areas farther south.

Figure 3 summarizes the structural interpretations for South Mountain near Caledonia Park, Pennsylvania. Figure 4 is an index map of recent areal geology investigations in the Blue Ridge north of Roanoke, Virginia.

ACKNOWLEDGMENTS

The writer is indebted to Dr. Robert Scholten, of The Pennsylvania State University, who gave freely of his time for instruction and assistance. His interest and direction throughout all phases of this study were invaluable. To Drs. D. Hoskins and S. Root of the Pennsylvania Topographic and Geologic Survey, the author expresses his gratitude for their interest and assistance particularly during the mapping phase of the investigation. Dr. Hoskins also critically read the manuscript.

Field meeting and discussions with Drs. Jacob Freedman, Carlyle Gray, Vinton Gwinn, Richard Nickelsen, and Arthur Socolow in the Caledonia Park area and at other localities in the Blue Ridge were extremely beneficial and sincerely appreciated.

Numerous discussions with various members of the faculty and the author's colleagues at The Pennsylvania State University are gratefully acknowledged.

Special mention is given to the Pennsylvania Topographic and Geologic Survey for supporting the field research, and The Pennsylvania State University, Department of Geology and Geophysics for graduate assistantships (1959-1964) and for underwriting the cost of petrographic thin sections.

PHYSIOGRAPHY

GENERAL STATEMENT

Topographically, the area is divisible into three longitudinal zones. Buchanan Valley, the eastern zone, and the ridges of South Mountain, the central zone, comprise nearly 70 per cent of the total area and represent the Blue Ridge. The third zone is the Cumberland Valley segment of the Valley and Ridge province.

The highest elevation, 2100 feet, occurs atop Big Pine Flat Ridge about five miles north of Caledonia Park. Just northeast of Scotland the elevation drops to 675 feet, the lowest in the area. Maximum relief is 1425 feet.

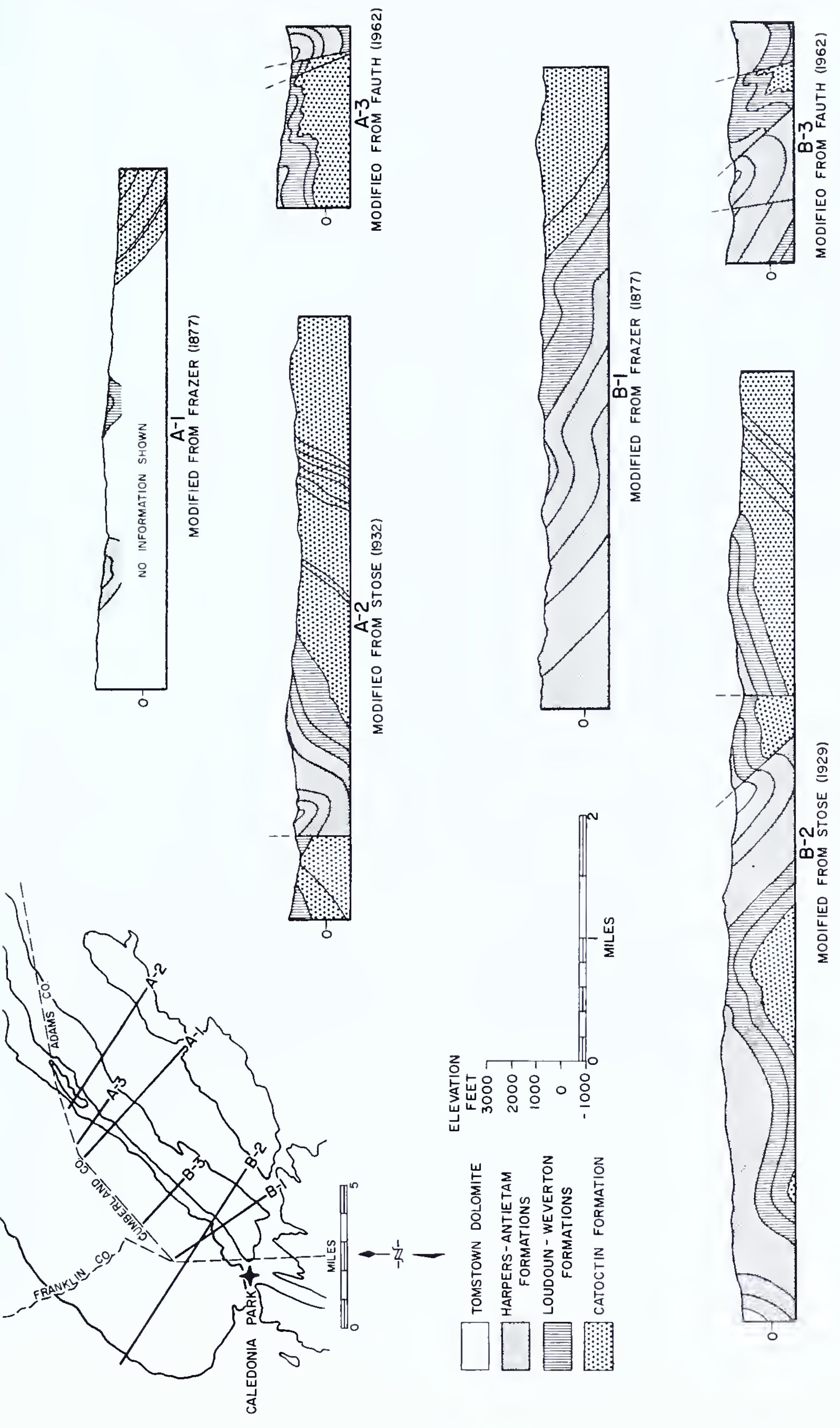

CALEDONIA PARK
FRANKLIN CO.
CUMBERLAND CO.
ADAMS CO.
A-1
A-2
A-3
B-1
B-2
B-3
N
MILES
0
5
TOMSTOWN DOLOMITE
HARPERS-ANTIETAM FORMATIONS
LOUDOUN-WEVERTON FORMATIONS
CATOCTIN FORMATION
ELEVATION FEET
3000
2000
1000
0
-1000
MILES
0
1
2
NO INFORMATION SHOWN
A-1
MODIFIED FROM FRAZER (1877)
A-2
MODIFIED FROM STOSE (1932)
A-3
MODIFIED FROM FAUTH (1962)
B-1
MODIFIED FROM FRAZER (1877)
B-2
MODIFIED FROM STOSE (1929)
B-3
MODIFIED FROM FAUTH (1962)

Figure 3. Evolution of structural interpretations of South Mountain, Pennsylvania.

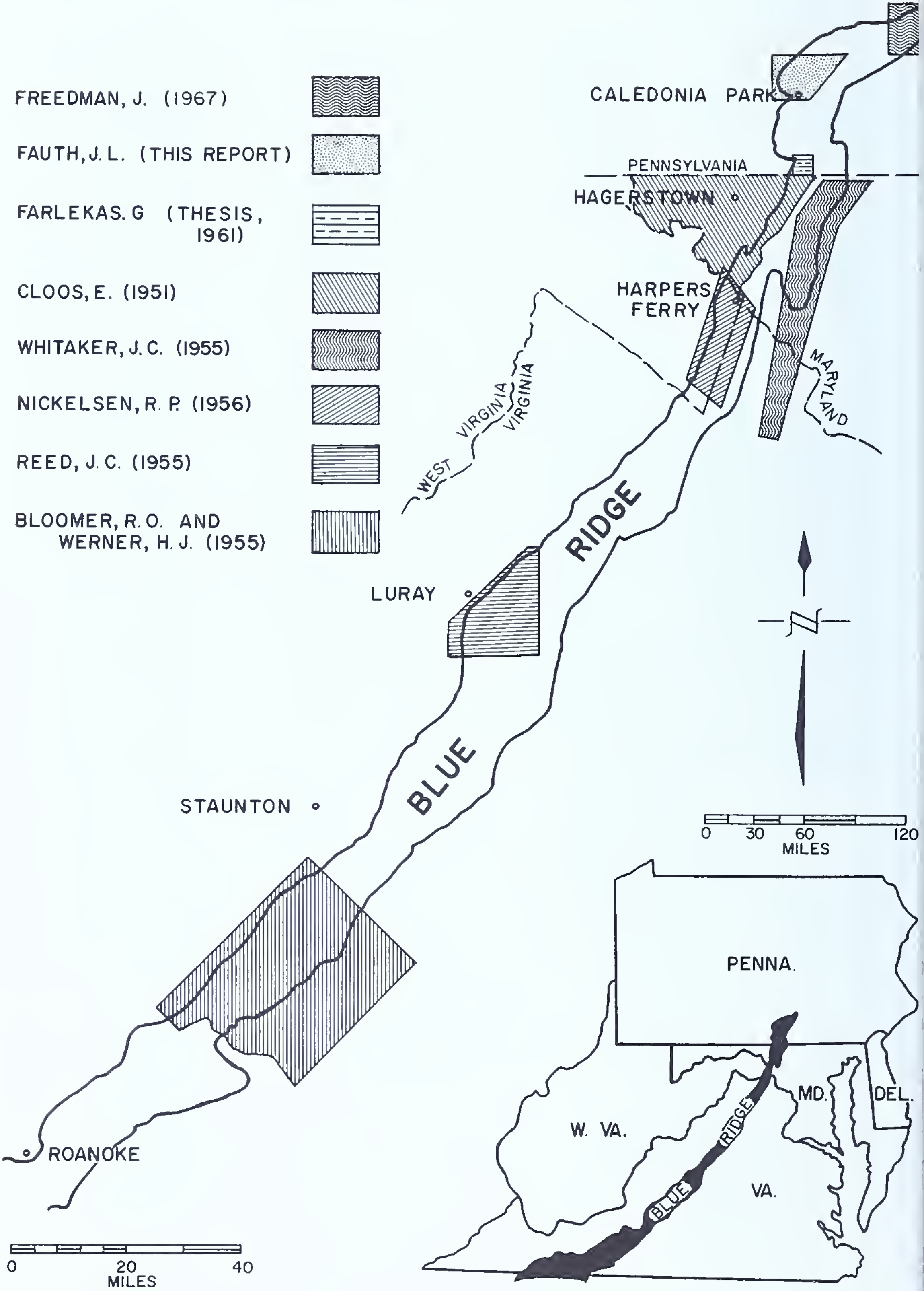


Figure 4. Location of some recent investigations in the Blue Ridge north of Roanoke, Virginia, from 1951 to 1966.

With the exception of the eastern part of the Cumberland Valley where a thick colluvial cover extends 8,000-10,000 feet out from the mountain front, the topographic trends and features tend to reflect the structural grain and stratigraphy of the area.

LANDFORMS

The valley of Conewago Creek, called Buchanan Valley, encloses about 10-12 square miles along the eastern margin of the area. The regional topographic slope declines eastward at 320-400 feet per mile. The surface form is moderately rolling with a maximum relief of 450 feet. Locally, relief rarely exceeds 200 feet except in the vicinity of major tributaries to Conewago Creek.

The series of northeast-trending ridges of South Mountain cover about 65 square miles and comprise the central topographic zone. A narrow transverse valley separates a northern group of ridges trending N45E and covering 80 per cent of this zone, and a southern group heading N25E.

Individual ridges in the central zone are separated by narrow, somewhat discontinuous valleys. Because the ridges have sub-rounded to flat-topped summits, and concordant summit levels between the elevations 1750 and 2100 feet, the western two-thirds of this zone has the form of a broad, moderately dissected, upland surface.

In the central zone slopes average 8°-15°. However, slopes of 20°-30° are not uncommon along the flanks of many of the deep, V-shaped stream valleys. Maximum relief is 1200 feet, and local relief is 450-650 feet. Lowlands in this zone are local and found only along Conococheague Creek near Caledonia Park.

The Cumberland Valley is a 20 square mile area of low, rolling topography west of South Mountain. Slopes are gentle almost everywhere ($<5°$), and in general decline westward at 100-150 feet per mile. Maximum relief is 325 feet while local relief in excess of a few tens of feet is uncommon.

DRAINAGE

The streams of the map area represent portions of two drainage systems. Conococheague Creek is in the drainage system of the Potomac River, whereas Mountain Creek and Conewago Creek flow into the Susquehanna River.

Conococheague Creek flows southwest in the east central part of the area from its source on East Big Flat Ridge to Caledonia Park, a distance of about six miles. Its gradient over this interval is 125-130 feet per mile as measured from topographic maps. At Caledonia Park, Conococheague Creek turns abruptly and flows eastward in a transverse valley developed along the trace of the Carbaugh-Marsh Creek fault. The creek continues in this direction up to the border of the map area 2½ miles northwest of Fayetteville. In the section west of the mountains, Conococheague Creek has developed an extensive floodplain whose width increases from 500 feet to 4000 feet near the western boundary, has a moderately anastomosing stream channel, and a gradient of 30 feet per mile.

Mountain Creek, near the northwest corner of the map, has a length of about three miles within the area. It flows in a moderately narrow but nearly flat valley in which small meander zones are locally well developed. Its gradient is about 45-50 feet per mile.

Conewago Creek heads immediately north of U.S. 30 near the eastern border of the mapped area and leaves the area southwest of Strasbaugh School. The average gradient of 100-110 feet per mile is over a distance of 1.5 miles.

Small streams which head in South Mountain and flow westward into the Cumberland Valley have sharp V-shaped valleys, and gradients between 300-400 feet per mile in the mountains. However, upon emerging onto the alluvial-colluvial apron west of the mountain front, the gradient drops off markedly (150-170 feet per mile), and only the larger streams (Mountain Run, Phillamen Run) are through-flowing into the Conococheague Creek.

STRATIGRAPHY

GENERAL STATEMENT

The core of the South Mountain anticlinorium is a Pre-cambrian granitic and gneissic complex termed the Injection Complex (Jonas and Stose, 1939, p. 575-580). The northernmost exposure of this unit is in Frederick County, Maryland where mica schist and hornblende diorite are intruded by granite and granodiorite.

Overlying the Injection Complex unconformably is the Swift Run Formation (Jonas and Stose, 1939, p. 585). This formation, which has a highly variable lithologic character, parallels the Blue Ridge as far north as southern Maryland. Near Harpers Ferry, West Virginia it consists of sericitic quartzites and phyllites with discontinuous lenses of marble near the base.

The Catoctin Formation is a thick sequence of volcanic rocks conformably overlying the Swift Run. In Pennsylvania the Catoctin Formation includes both rhyolites and basalts. Farther south the rhyolites become subordinate to the basalts, eventually disappearing in southern Maryland.

A thick section of clastic rocks overlie the volcanic rocks. These rocks, called the Chilhowee Group, include the Loudoun Formation, Weverton Quartzite, Harpers Formation, and Antietam Quartzite. This group is assigned an Early Cambrian age because its upper part (Antietam Quartzite) contains Lower Cambrian fossils.

The Tomstown Formation, which overlies the Chilhowee Group, represents the basal portion of the extensive Cambro-Ordovician carbonate sequence. The Waynesboro, Elbrook, and Conococheague Formations plus the base of the Beekmantown Group are the additional representatives of this sequence exposed in the area studied. Overlying the carbonates are unconsolidated Cenozoic alluvial and colluvial deposits.

Table 1 summarizes the complete stratigraphic sequence in the study area, and Figure 5 provides a correlation of the Caledonia Park section with equivalent units to the south.

PRECAMBRIAN ROCKS

Catoctin Formation

The Catoctin volcanics are the only Precambrian rocks exposed in the Caledonia Park area. Geiger and Keith (1891, plate IV) named the Catoctin Formation for exposures in Catoctin Mountain, northern Virginia. Williams (1892, p. 482-496) and Bascom (1896) were the first to recognize the volcanic character of this unit. Reed (1955, p. 871-896) recently studied this volcanic sequence near Luray, Virginia. There excellent exposures in the upper limb of the Blue Ridge anticline, where deformation has been minimal, afford an opportunity to determine the thickness and original nature of the lavas from which the Catoctin greenstone was derived.

The typical greenstone in the Luray, Virginia area is a greenish, fine-grained, altered basalt which overlies a basal unit of arkoses, graywackes, and conglomerates. Locally, yellowish-green epidosite and sericitic phyllites occur interbedded with the volcanics (Reed, 1955, p. 879).

Stose and Stose (1946, p. 20-24) describe green metabasalt, hornblende schist, tuffaceous beds, blue andesite, and gray to blue, porphyritic aporhyolite from the Catoctin Formation of northern Maryland.

Starting at the southern boundary of Pennsylvania the Catoctin Formation occurs about six miles east of Waynesboro, and extends as a belt 2-3 miles wide along the eastern flank of South Mountain as far north as Dillsburg. Stose (1932, p. 28-30) describes several varieties of altered rhyolites and rhyolite breccias from the Catoctin in this area in addition to metabasalt, chlorite and sericite schists, and epidosite.

TABLE 1. Generalized stratigraphic column in the Caledonia Park Area. Thicknesses are in feet.

CRETACEOUS (?), TERTIARY - QUATERNARY

Alluvium

Unconsolidated silt, sand, and gravel. Thickness: 0-20+.

Terrace Gravels (Alluvial-Colluvial Deposits)

Unconsolidated deposits of white, gray, and bluish sandstone and quartzite pebbles, gravel, and boulders generally with a clayey matrix. Thickness: 0-75+.

ORDOVICIAN

Rockdale Run Formation

Interbedded limestone and dolomite; algal and fragmental limestone mainly in lower two-thirds, laminated dolomite and dolomitic limestone in upper part. Thickness: 1900-2500.

Stonehenge Formation

Basal thin-bedded sandy and conglomeratic limestones overlain by mechanical limestone; coarser in upper half with tabular limestone fragments 2-3 inches in diameter. Thickness: 800.

CAMBRIAN

Conococheague Formation

Cyclic limestone with prominent laminations and bands of silty dolomite, and subordinate oölitic and fragmental limestone above a basal calcareous sandstone; pink, finely crystalline limestone and black chert near the top. Thickness: 1800-2000.

Elbrook Formation

Yellow-orange-weathering, greenish and gray shales and finely-laminated, shaly limestone in lower half; gray limestone above with calcareous sandstone, oolite, and laminated limestone near top. Thickness: 3000.

Waynesboro Formation

Bluish limestone and dolomite above basal siliceous gray limestone; interbedded greenish siltstones and shales with prominent purple shales in upper portion. Thickness: 1250.

Tomstown Dolomite

Gray dolomite with interbeds of white limestone; thin bedded blue limestone near middle, sericitic schist at base. Thickness: 1000.

Chilowee Group

Antietam Quartzite

Medium-to coarse-grained, white to grayish protoquartzite and quartzite; *Skolithos* present. Thickness: 700-900.

TABLE 1. Generalized stratigraphic column in the Caledonia Park Area. Thicknesses are in feet. (Cont.)

Harpers Formation

Upper Harpers

Green to greenish-gray, fine-grained quartzose graywacke and graywacke. Thickness: 300-500.

Montalto Quartzite Member

Upper unit: white to light-gray, medium-to coarse-grained protoquartzite and quartzite with blue quartzite zone near middle; *Skolithos* present. Thickness: 700-800

Lower unit: gray, medium-grained protoquartzite and quartzite, typically thin-bedded and cross-bedded; *Skolithos* present. Thickness: 900-1100.

Weverton Quartzite

Interbedded series of grayish-green and purplish protoquartzite and quartzose graywacke with irregularly distributed intervals of thin graywacke conglomerate, quartzite, and phyllite. Thickness: 1200-1400.

Loudoun Formation

Conglomerate Member

Interbedded grayish-green and bluish-gray graywacke or quartzose graywacke and a quartz-phyllite-rhyolite pebble conglomerate. Thickness: 0-200.

Phyllite Member

Predominantly reddish-purple and grayish-blue phyllites commonly exhibiting greenish or white blebs or streaks. Thickness: 150-250.

PRECAMBRIAN

Catoctin Formation (stratigraphic sequence not known)

Metabasalt

Greenish-gray, fine-grained metabasalt; infrequently coarse-grained, porphyritic, or amygdaloidal.

Epidosite

Fine-to medium-grained, granular, yellow-green epidosite with a few amygdaloidal zones.

Greenstone

Metabasalt and epidosite undifferentiated.

Phyllite

Light-greenish to green-gray phyllite; rarely banded in combinations of bluish, pink, and green.

Bluish metarhyolite

Bluish and reddish, uniformly aphanitic or porphyritic metarhyolite; locally banded or brecciated, rarely vesicular.

Mottled metarhyolite

Black and brownish-gray, porphyritic metarhyolite with phenocrysts of both quartz and feldspar.

Red porphyritic metarhyolite

Red to lavender metarhyolite containing phenocrysts of both quartz and feldspar.

Metarhyolite breccia

Breccia composed of gray to bluish metarhyolite fragments in a reddish matrix.

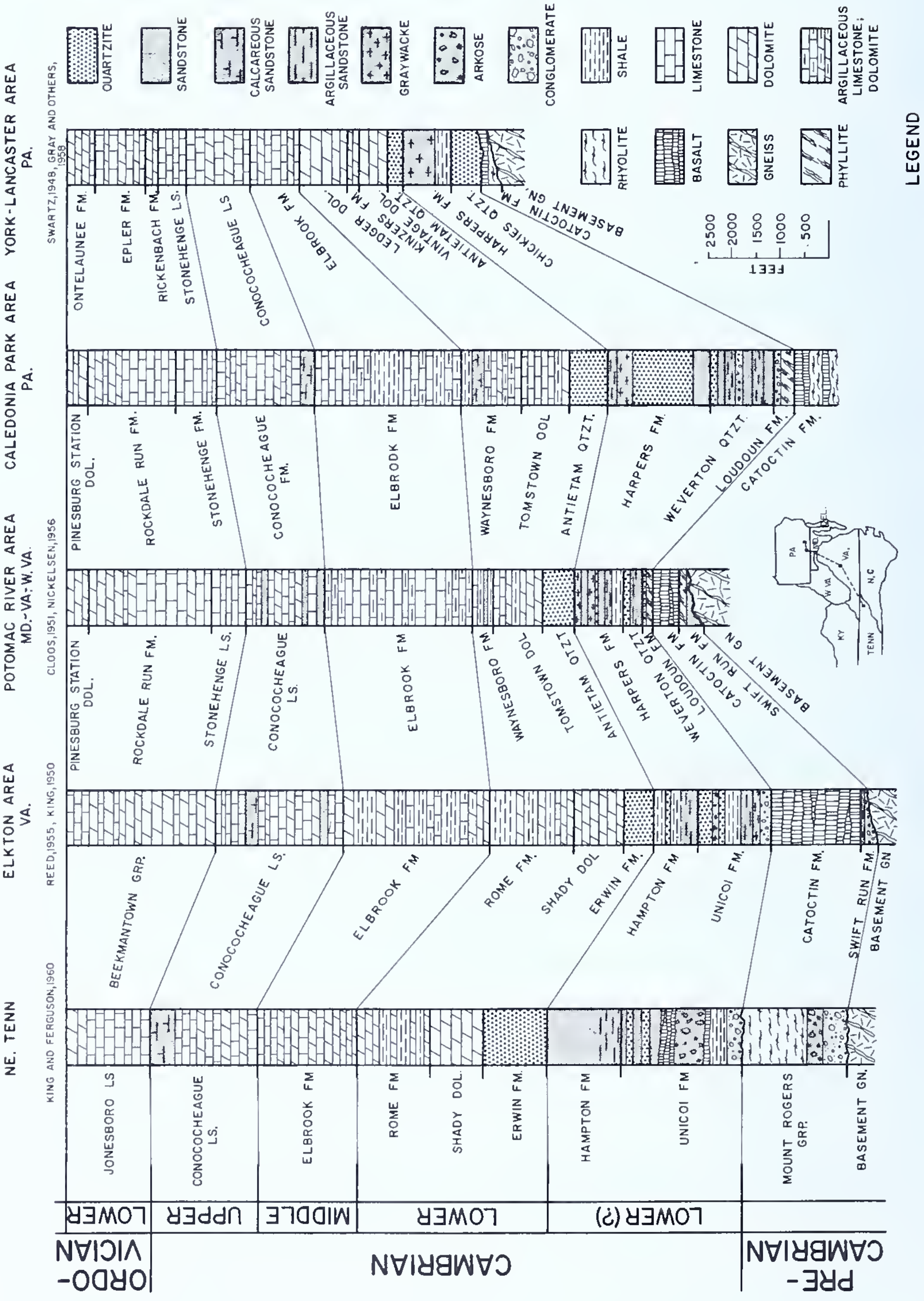


Figure 5. Stratigraphy of the Blue Ridge from northeastern Tennessee to Pennsylvania.

The diverse lithologic character of the Catoctin Formation is not reflected in previous geologic maps of the area because only four lithologic units were distinguished (Stose, 1932; Stose and Bascom, 1929). The map pattern thus determined does not outline structures similar to those mapped in the overlying metasediments.

In an attempt to more clearly delineate the structures formed in the volcanic sequence and to gain added information on the complex stratigraphy of this unit, the writer subdivided the three major lithologic groups of the Catoctin formation (greenstones, metarhyolites, and phyllites) into eight mappable lithologic units.

Metabasalt

The typical metabasalt of the area is a greenish-gray (5 G 6/1) to dark-greenish-gray (5 G 4/1), uniformly aphanitic to fine-grained rock. Locally, the rock has a gray to bluish color or is coarse-grained and shows small (1-4 mm), dusky-green (5 G 3/2) spots in a greenish-gray groundmass.

Amygdaloidal and porphyritic metabasalts also occur. The amygdaloidal rock contains colorless to milky white, roughly circular amygdules 1.0-6.0 mm in diameter. Either feldspar, epidote, quartz, or chlorite usually fill the amygdules. Locally two of these minerals may fill an amygdule.

Laths of feldspar up to 25 mm long characterize the porphyritic variety. The phenocrysts, which average 5-8 mm, are readily visible in the weathered rock because of their bone white color.

The regional cleavage and joints are the dominant planar structures in outcrop. Primary flow structure, columnar jointing, and pillow structures were not recognized in the metabasalt of the area.

Veins of milky quartz, sometimes in association with epidote or a fibrous amphibole, commonly cut the metabasalt. The metabasalt commonly includes lenses and irregular pods of granular epidote and quartz. When most of the typical metabasalt is replaced by epidote and quartz, a tough granular rock, epidosite, is formed.

The metabasalt occurs in the central and northern parts of Buchanan Valley, and in a limited area along Rocky Mountain Creek, about two miles south of Caledonia Park.

Exposures of the even-grained metabasalt are located in the vicinity of Boyd Run, 0.25-0.4 mile north of BM 1261 on the Shippensburg-Arendtsville road. The rock outcrops to the south of the stream just east of the macadam road, and again on the north side of the stream between the elevations 1240-1380 feet.

The amygdaloidal rocks are poorly exposed, but outcrop is found along and west of Route 234, ½-¾ mile northeast of Strasbaugh School.

Only scattered blocks of the porphyritic metabasalt occurs; none has been located in outcrop. This float is seen along the south side of the dirt road 0.7 mile southeast of Strasbaugh School, between the elevations of 1320-1400 feet.

The coarse-grained metabasalt is restricted to a few localities in the valley of Rocky Mountain Creek south of Caledonia Park, and near Carbaugh Run between Snaggy River and Mt. Newman.

Epidosite

Quartz and epidote combine to form epidosite, a uniform, fine- to medium-grained, granular rock whose dusky-yellow-green (5 GY 5/2) to moderate-yellow-green (5 GY 7/4) color is distinctive.

An amygdaloidal variety in which the amygdules, 2-10 mm in diameter, are filled with white and light-gray quartz commonly rimmed by epidote is also recognized. The groundmass usually has a yellow-green, or mottled, moderate-yellow-green (5 GY 7/4) and pale-blue-green (5 BG 7/2) color.

The epidosite, in the form of large, irregular pods, lenses, veins, and oval knots, occurs within the metabasalt. The contact with the metabasalt may either be sharp or gradational. In many places the veins and lenticular bodies of this rock occur along the regional cleavage. Closely spaced fractures within the epidosite and normal to the regional cleavage may be filled by quartz or quartz associated with an asbestiform amphibole. In contrast large pods and zones of this quartz-epidote rock, such as observed in a greenstone quarry on Pine Mountain near Gladhill, Pennsylvania, outside of the study area, show no visible relation to structures developed in the surrounding rocks.

In the study area, epidosite has been observed in outcrop only as small nodular masses, lenses, and veins enclosed by metabasalt. The map pattern showing the distribution of epidosite has been determined entirely on the basis of float. A ratio of epidosite to metabasalt float of at least 4:1 was the basis for distinguishing and mapping this unit.

Examples of typical epidosite are numerous along the Shippensburg-Arendtsville Road, between the elevations 1320-1360 feet, approximately 2½ miles northeast of Strasbaugh School.

Greenstone

This map unit designates areas of metabasalt and epidosite float, and is used when the concentration of either metabasalt or epidosite in the float falls below about 80 percent.

Bluish Metarhyolite

This unit is dominated by metarhyolite and metarhyolite breccia of moderate-bluish-gray to grayish-blue color. Banded metarhyolite and piedmontite-bearing metarhyolite and metarhyolite breccia of light-grayish-red (5 R 4/2) to grayish-red-purple (5 RP 4/2) color constitute a major secondary group.

Locally the rocks of this unit are purple, green, brownish or grayish in color. On weathering, the metarhyolites usually show lighter shades of the above colors.

Although the bluish and reddish metarhyolite groups each have a considerable areal extent, they are gradational one into the other and no consistent or workable field criteria could be established to distinguish between them throughout the area.

Porphyritic and even-grained, aphanitic textures are the most common; these rocks may also exhibit flow structure and banding. In one locality east of Rock Mountain Creek, approximately two miles south of Caledonia Park, a porphyritic metarhyolite was collected which proved to be vesicular upon microscopic examination. Banding occurs in some specimens.

All of the rhyolite lithologies described previously occur as fragments in the breccia except for the vesicular rhyolite. The flow-banded and porphyritic types are most abundant.

In general, the metarhyolite is a hard, dense rock which tends to break with a conchoidal fracture, or along closely spaced joints. The cleavage developed in these rocks is most evident in weathered exposures.

White, pink, or orange laths of feldspar occurring alone or with almond-shaped, grayish quartz phenocrysts, 1-2 mm in diameter, are characteristic of the porphyritic rocks. The estimated average length of the feldspar laths is 3-4 mm and rarely exceeds 7 mm. Phenocrysts constitute 5-20 percent of this metarhyolite variety.

The banded metarhyolites are characterized by a light-grayish-red (5 R 4/2), grayish-red-purple (5 RP 4/2), or grayish-blue color with thin (1.0 mm), dusky-red (5 R 3/4) to grayish-red-purple bands or streaks occurring at variable intervals. A few of these metarhyolites consist of alternating light-gray, pale-red-purple (5 RP 6/2), and grayish-purple bands 0.5-4.0 mm thick.

The typical breccia is light bluish gray (5 B 7/1) to dusky blue (5 PB 3/2). It contains generally angular metarhyolite fragments from a few millimeters up to 25 cm or more in diameter in a darker bluish-gray to purplish matrix. These blocks often exhibit a crude dimensional alignment in the plane of the regional cleavage.

Piedmontite, commonly as fibrous crystals, occurs in veins, irregular patches, and gash fractures in the non-banded, reddish metarhyolites. A breccia containing pale-red-purple to pale-red (5 R 6/2) fragments, some of which are porphyritic, has a matrix rich in quartz and piedmontite. Piedmontite has also been found, but in much smaller quantities, in a uniformly aphanitic, grayish-purple to grayish-red-purple rock.

The best exposures of this bluish metarhyolite and metarhyolite breccia unit are on both sides of the dirt road about 3000 feet southwest of Strasbaugh School, on the east side of Carbaugh Run between the elevation of 1120-1140 feet, and along Rocky Mountain Creek, about one mile south of Caledonia Park.

Good exposures of the banded metarhyolite are on the golf course adjacent to Carbaugh Run, 3000 feet east southeast of Caledonia Park, and east of the macadam road between the elevations 1220-1240 feet, 2.4 miles southeast of Wenksville. Near the eastern margin of the map area, just northeast of the creek approximately 1.9 miles northeast of Strasbaugh School, one of the piedmontite-bearing rocks is exposed. The piedmontite-bearing breccia is found only in float, and occurs just west of the macadam road 2.2 miles southeast of Wenksville between the elevations 1210 and 1260 feet.

Mottled Metarhyolite

Thin black streaks and bands set in bluish-gray and yellow-gray to light-brownish, porphyritic groundmass characterize this unit and give the rock a mottled appearance. Laths of white or pinkish-gray feldspar, 0.5 - 6.0 mm long, and medium- to dark-gray quartz occurs as phenocrysts. The quartz grains are about 1.0 - 2.0 mm in diameter.

In exposures near the District Road just below the summit of Snaggy Ridge (elevation 1540-1600 feet), the black streaks in the rock have a marked orientation. These color streaks, which probably represent a primary flow or bedding structure, correspond to an *a* lineation in the plane of the cleavage.

This lithology is exposed only in the southeastern corner of the area. It is prominent along the crest and eastern slopes of Snaggy Ridge and the flanks of Mt. Newman.

Metarhyolite Breccia

A breccia of pale-blue to moderate-grayish-blue rhyolite fragments enclosed by a moderate- to dark-red matrix describes this unit. Pink and black banded fragments and maroon jasper (?) appear locally. The rock weathers light gray to bluish gray, and red purple.

The breccia fragments are equant and range in size from 2.0 - 50.0 mm. Although most are uniformly aphanitic, some fragments have a porphyritic texture and flow structure. Reddish-brown to orange feldspar laths commonly less than 5 mm long are the phenocrysts in the porphyritic types.

This rock is restricted to a narrow, short, arcuate zone just south and west of Strasbaugh School. Accessible, but dubious, exposures occur along the west side of Route 234, about 1000 feet southwest of Strasbaugh School.

Red Porphyritic Metarhyolite

A moderate-red to grayish-red, or pale-red-purple rock which weathers moderate dusky red (5 R 4/4) to very pale red is characteristic of this rhyolite unit. Less commonly, the rock is colored grayish red purple (5 RP 4/2).

The distinguishing feature of this rock is the occurrence of numerous phenocrysts of white feldspar and deep-reddish colored quartz. The average length of the feldspar is 2.0 - 4.0 mm, whereas the quartz averages about 2.0 mm.

No primary structures are recognized in exposures of this rock; however, cleavage is usually well developed.

Outcrops of this metarhyolite unit are uncommon; the only area of exposure is located immediately northwest of the junction of two dirt roads, 1.25 miles north of Strasbaugh School. Abundant float of this unit can be observed in the immediate vicinity of Strasbaugh School and nearby localities.

Phyllites

Isolated, mappable zones of phyllite are associated with the metarhyolites and metabasalts. These rocks are basically pale green to light greenish gray and weather greenish gray to yellowish gray. Less commonly, the rock is light grayish blue with thin, pale-pink bands, light green with very-light-gray bands, or nearly white with thin, irregular pink bands. The grayish-blue rock has a mottled purplish and orange appearance on the weathered surface.

Cleavage is the most conspicuous planar structure in these rocks. One cleavage plane is developed in all of the phyllites, and locally, two distinct directions are prominent.

Many of the phyllite zones are marked by the occurrence of prominent veins and float of milky white quartz. Some of the phyllites themselves are intruded by small veins of quartz, usually along the older foliation, or are partially silicified.

Accessible exposures of the phyllite unit occur at approximately the elevation 1200-1240 feet about 300 feet west of Route 234, 7000 feet southwest of Strasbaugh School. Other exposures are located along the west side of the macadam road, between the elevations 1160-1210 feet, 2000 feet southwest of Wenksville.

Stratigraphic Relations

The relative age of the Catoctin metarhyolites and metabasalts is a question that cannot be satisfactorily answered for the area studied because of poor exposure and the absence of useable primary structures in these rocks.

Contacts between the two major lithologic members of this formation have not been observed except for that of the epidosite and metabasalt. Contact relations between rhyolite and basalt are restricted to five localities in Pennsylvania as noted in the literature. In four localities near Monterry, Pennsylvania, outside the study area, the rhyolites are interpreted to be overlain by the basalts (Bascom, 1896, p. 29). However, a few miles north of these localities, G. H. Williams recognized the reverse situation (Bascom, 1896, p. 28).

Stose and Stose (1946, p. 26-27) suggests that near Pine Knob, Maryland the rhyolites and associated pyroclastics occupy a syncline in the basaltic terrane and therefore are younger than the basalts. However, in Pennsylvania, Stose (1929, p. 4) had concluded that the metabasalts were the younger rock.

Whitaker (1955, p. 439) recognizes rhyolite tuff in gradational contact with basalt above and below near Pen Mar, Pennsylvania, about 12 miles south of Caledonia Park, and suggests an interbedded relationship.

In the absence of exposures showing distinct contact phenomena and without better structural control, the evidence regarding the relative age of the rocks of the Catoctin Formation is ambiguous, and thus, the stratigraphy of this unit remains unsettled.

The contact between the Catoctin volcanics and the overlying sediments of the Chilhowee Group is important because:

(1) This contact is interpreted as marking the base of the Cambrian by some authors (King, 1949; Stose, 1932), whereas others assign an Early Cambrian age to the volcanics (Cloos, 1951; Bloomer and Bloomer, 1947) or a Precambrian(?) age to Catoctin plus all rocks below the fossiliferous Antietam Quartzite (Howell and others, 1944).

(2) The uncommon exposure of the contact zone throughout the Blue Ridge area makes it difficult to ascertain the true physical character of the boundary, whether conformable or unconformable.

The question of the location of the base of the Cambrian has been analyzed in detail by King (1949, p. 512-530, 622-645) for the Southern Appalachians, and has been discussed in most of the literature on the geology of the central and northern Blue Ridge.

The contact between the volcanics and Chilhowee rocks is not exposed in the area studied. However, the geologic map (Plate 1) shows both metarhyolite and metabasalt in contact with the Loudoun Formation, the basal member of the Chilhowee Group. This geometric relationship can be satisfactorily explained in three ways:

(1) The contact marks an angular unconformity due to folding of the volcanics before Chilhowee sedimentation.

(2) There is structural conformity across the contact but the Chilhowee sediments are deposited on a topographic surface of moderate erosional relief such that both rhyolites and basalt were exposed; no intervening period of folding is implied.

(3) The boundary is a conformable or essentially conformable contact — any erosional period was of low intensity and short-lived, and the map pattern primarily reflects the discontinuous lateral and interbedded nature of the rhyolites and basalts.

Reed (1955, p. 878) and Bloomer and Werner (1955, p. 598) report that no angular unconformity exists in the central Virginia area in exposures showing the contact between the Loudoun Formation and the Catoctin. Instead, the authors report a conformable intertonguing and interbedded relationship. However, King (1949, p. 526-528, 635) does report an angular unconformity from the Elkton area of Virginia.

Additional evidence against a structural discordance is provided by the fact that the intensity of deformation and orientation of structural elements (i.e., cleavage, lineations) pass without change from the volcanics into the overlying sediments throughout the Blue Ridge (Cloos, 1951, p. 25-28). This relationship is corroborated by the present study.

Reed (1955, p. 878-879, 891-893) recognizes a purple slate between and conformable to both the Catoctin Formation and the Chilhowee Group. This rock, described as a purplish, intensely cleaved, and containing green chloritic blebs elongated down the cleavage, is found to exhibit a relic basaltic texture under the microscope. In addition, the green blebs appear to represent either amygdules or chloritized ferromagnesian phenocrysts (Reed, 1955, p. 893). Field relations and microscopic character led Reed (1955, p. 893) to suggest that this lithology represents a metamorphosed saprolith developed at the top of the Catoctin Formation prior to the deposition of the overlying sediments. Furcron and Woodward (1936, p. 50-51) describe a similar rock conformable with the base of the Loudoun, but interpreted it as a lava flow of andesitic (?) character. Similar rocks at this horizon have also been noted by King (1950), Nickelsen (1956), Whitaker (1955), and Fauth (1962).

A similar phyllite is present along much of Piney Mountain east of Caledonia Park. Although all relic structures have been destroyed by intense shearing and recrystallization, the lithologic appearance and stratigraphic position of this phyllite is strikingly similar to that reported by Reed.

Cloos (1951) reports that at many localities in Washington County, Maryland there are gradational changes from the Catoctin into the Loudoun. "Sand grains appear in well-bedded volcanics below the Loudoun and tuffaceous purple bands are interbedded with the Loudoun . . . Downward and upward the change is gradational and not abrupt as should be expected if an unconformity were present". (Cloos, 1951, p. 27).

On the basis of the above information and lacking other direct evidence from the study area, it appears most probable that regionally the Catoctin-Chilhowee boundary does not represent a lengthy period of intensive erosion, and it certainly is not an angular unconformity. Thus, the map pattern of the Catoctin-Chilhowee zone in the Caledonia area probably reflects the deposition of the Loudoun sediments over a terrain of low relief formed mostly by the irregular distribution and discontinuous character of a series of relatively thin individual rhyolite and basalt flows.

Thickness of the Catoctin Formation

Thickness estimates for the Catoctin Formation in areas where the unit is composed wholly of metabasalt range from 0-50 feet at the Potomac River near Harpers Ferry, and 500-1000 feet five miles northwest of Purcellville, Virginia (Nickelsen, 1956, p. 245), to about 1800 feet near Luray, Virginia (Reed, 1955, p. 883).

Because of poor exposures, inadequate knowledge of the structure, and the common termination of the formation on the east by Triassic border faults, reliable estimates of thickness are not available where the volcanic sequence consists of both rhyolites and basalts.

In Adams County, Pennsylvania, Stose (1932, p. 29) assigns a thickness of 1000± feet each to the rhyolites and basalts. Assuming structural complexity in the volcanics similar to that in the overlying metasediments, a total thickness for the unit of 1500-2000 feet is probably a minimum figure.

CAMBRIAN ROCKS

The Chilhowee Group, a series of coarse clastic rocks, forms the lower half of the Cambrian section in the Caledonia Park area. The general petrographic character of this group is given in Figure 6 and Table 2. Percentage estimates are based on point count data and visual comparison with charts by Terry and Chilingar (1955, p. 229-234). Sorting is described on the basis of the number of Wentworth size grades including 80 percent of the detrital grains as follows: one Wentworth size grade, very good; 2-3, good; 4-5, moderate; 6-7, fair; more than seven size grades, poor. A section of carbonate rocks, dominated by limestones, overlies the Chilhowee Group and comprises the remainder of the Cambrian System.

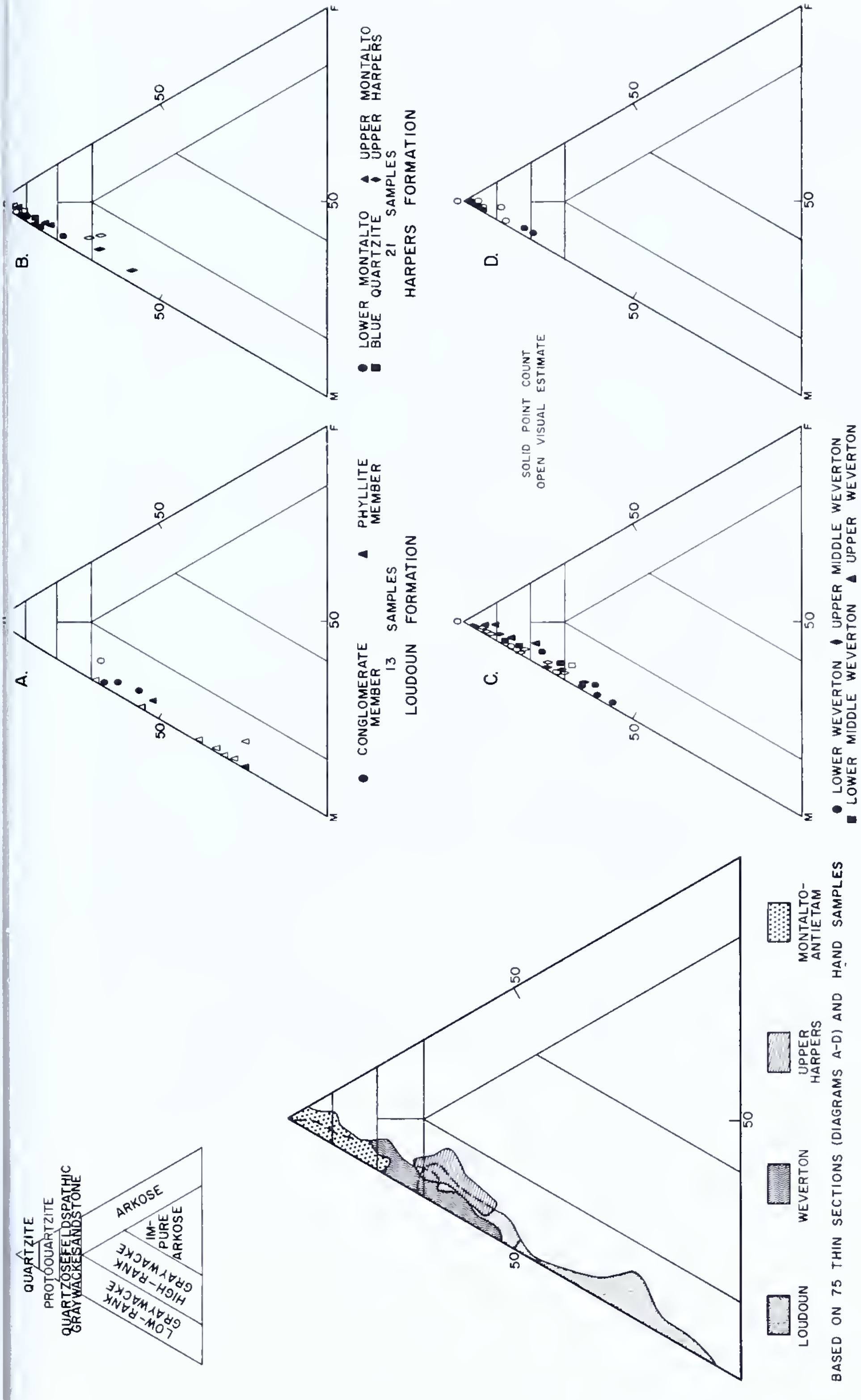


Figure 6. Composition of the Chilhowee Group in south-central Pennsylvania.

TABLE 2. Summary of sedimentary characteristics in the Chilhowee Group.

Formation	*Quartz*		*Feldspar*	*Sorting*	*Roundness*	*Non-Opaque Heavy Minerals**
	(Percent)	*(Percent)*	*(Type)**			
Antietam	Range 78-99 Mean 91	Range 0-3 Mean <1	Plagioclase Microcline	Good	Subrounded to rounded	TOURMALINE, ZIRCON, Rutile
Harpers (upper)	Range 60-71 Mean 64	Range 3-6 Mean 4	MICROCLINE Plagioclase	Moderate	Subrounded	TOURMALINE, Zircon
Montalto (upper)	Range 82-98 Mean 89	Range 0-1 Mean<<1	MICROCLINE Plagioclase	Good	Subrounded	Tourmaline, Zircon
Montalto (lower)	Range 82-97 Mean 89	Range 0-2 Mean 1	MICROCLINE Plagioclase	Good	Subrounded	Tourmaline, Zircon
Weverton (upper members)	Range 65-98 Mean 82	Range 0-4 Mean 2	MICROCLINE Plagioclase	Moderate to good	Subrounded to subangular	TOURMALINE, EPIDOTE, Zircon
Weverton (lower members)	Range 56-88 Mean 71	Range 0-3 Mean <1	PLAGIOCLASE Microcline	Moderate	Subrounded to subangular	TOURMALINE, EPIDOTE, Sphene, Zircon
Loudoun (conglomerate)	Range 58-69 Mean 66	Range 0-4 Mean 2	PLAGIOCLASE Microcline	Fair	Subrounded to subangular	SPHENE, EPIDOTE Zircon, Tourmaline

*Most abundant minerals capitalized.

Loudoun Formation

Name

The Loudoun Formation, the basal unit of the Chilhowee Group, was named by Keith (1893, p. 324) for exposures along Catoctin Mountain, Loudoun, Virginia.

Lithology

The Loudoun was originally described as a sequence of sandstones, conglomerates, and limestones (Keith, 1893, p. 324). This description was modified by Stose because certain lithologies mapped as Loudoun by Keith were, in fact, part of the older but lithologically similar Swift Run Formation (Stose and Stose, 1946, p. 19-20). Subsequently, however, the Loudoun as mapped by Stose was found to include, in one area, parts of the lower Harpers Formation (Cloos, 1951, p. 29) and elsewhere the lower Harpers and upper Weverton Quartzite (Whitaker, 1955, p. 441).

In the Caledonia area, Stose (1932, p. 42) described this unit as a sequence of "fine-grained, purplish gray, soft, sericitic schist or phyllite much of which is blotched by fragments of similar material, probably originally pebbles", overlain by "harder, schistose arkosic conglomerate and sandstone".

The writer recognizes two lithologic members, a basal phyllite and an upper conglomerate member. Exposures, particularly of the phyllite member, are limited. However, the two units are traceable on the basis of topographic expression and float.

Most of the phyllites are pale red purple to grayish red purple, medium gray, or grayish blue to moderate dusky blue (5 PB 3/2). Light-gray phyllite and slate, and banded red-purple phyllite to quartzose phyllite occur locally.

Conspicuous ivory or green colored blebs, streaks, and lenses, or pale-red (5 R 6/2) to dusky-blue phyllitic fragments one half to three inches long are common in these rocks. All are elongate and flattened in the plane of the major cleavage. The patches occur in rocks whose lithology and stratigraphic position is similar to the volcanic slate of Reed (1955, p. 291-293).

The contact between the phyllite at the base of the Loudoun Formation and the Catoctin Formation is not exposed in the area studied. It is drawn on the basis of the first occurrence of phyllite float above the highest observed volcanic fragments.

The upper member is a polymictic conglomeratic unit. Rounded one-half to two inch pebbles of gray and pink quartz, dusky-blue to red-purple phyllite, and grayish-blue to purplish rhyolite characterize this rock. Basalt and granite pebbles were not recognized in the rock. The rock has a grayish-blue to light-greenish-gray graywacke matrix. The matrix has well-developed cleavage in which the phyllite rock fragments are elongated and oriented downdip.

Grayish-blue and yellowish-gray weathering quartzose phyllites are interbedded with the conglomerate. Beds in this member usually are from one-half to four feet thick with cross-bedding occurring in a few of the phyllitic zones (Plate 2, Figs. 1, 2).

The conglomerate is a lens-shaped lithologic body. Where the conglomeratic member is present, the contact with the overlying Weverton Quartzite is gradational over approximately a 50 foot interval by a progressive decrease in the coarseness, thickness, and frequency of the conglomeratic beds.

Where the conglomerate member is absent, there appears to be an interbedded contact zone between the phyllites of the lower member and the more quartz-rich phyllites and graywackes of the basal Weverton.

The general sedimentary character of the Loudoun Formation is summarized in Figure 6 and Table 2.

Thickness

Because of the lack of exposure, intensity of deformation, and the nearly universal obliteration of bedding by cleavage, thickness of the phyllite member of the Loudoun is only an estimate. Nickelsen (1956, p. 247) suggests a thickness of 50-100 feet near Harpers Ferry, and Whitaker (1955a, p. 442) a maximum figure of 100 feet of Catoctin Mountain. Based on mapped outcrop width along the base of Piney Mountain at localities where the overlying Weverton dips steeply, the probable thickness of this unit is not less than 150-250 feet.

The conglomerate member varies in thickness from 0-200 feet. The mapped width of exposure of this member along the power line about 0.6 mile northeast of the Shippensburg Road where the conglomerate is nearly vertical, suggests the maximum thickness estimate.

A total thickness of 150-450 feet is thus estimated for the Loudoun Formation in the area studied.

Distribution

The Loudoun occupies linear belts near the base of the southeastern slope of Piney Mountain and East Big Flat Ridge (Plate 1).

Along Piney Mountain the phyllite member is continuous, but the conglomerate member is locally absent northeast of its first appearance, three miles north of U.S. 30. At the south end of Piney Mountain, there is no evidence to support the mapping of the Loudoun as shown by Stose (1929).

Both members of the Loudoun occur on East Big Flat Ridge, but only in the northern half of the area (Plate 1). The extension of these units almost to the southern end of the ridge by Stose (1929) is invalid, as the rocks exposed at these localities are quartzose phyllites of the overlying Weverton. The same applies to a mapped occurrence of Loudoun along Birch Run. Loudoun is likewise absent in outcrop and float along the eastern slope of Rocky Mountain, although Stose (1929) shows it there. Because the conglomeratic member is exposed at the end of Rocky Mountain, about five miles south of the map area, it is possible that the Loudoun is present at the base of Rocky Mountain but covered by the abundant Weverton float.

Exposures of the phyllites are rare. Gray phyllite crops out at elevation 1700-1740 feet just east of the Piney Mountain road, about 6000 feet southwest of its intersection with the Miltonberger Road. Abundant float of this member occurs along the Piney Mountain road between elevations 1700 and 1760 feet, and gray slate is exposed in an abandoned quarry west of Wenksville.

The conglomerate member is well exposed on Piney Mountain in the vicinity of the power line just north of the Shippensburg Road at the elevation of 1400-1500 feet.

Age and Correlation

No fossils have been found in the Loudoun Formation. It is generally assigned an Early Cambrian (?) age on the basis of its conformable stratigraphic and structural relationship to the overlying members of the Chilhowee Group, the uppermost part of which contains a Lower Cambrian fauna.

The formation was traced into the South Mountain area from its exposures in the type area by Stose (1929, 1932,) and Stose and Stose (1946). The writer has examined portions of the Loudoun at several localities south of the map area, notably, in the vicinity of Harpers Ferry, east of Waynesboro, Pennsylvania, on Route 16, and near South Mountain, Pennsylvania and concludes that the Loudoun of the Caledonia area is the same as in the type area.

Weverton Quartzite

Name

The Weverton Quartzite was named by Keith (1893, p. 329) for exposures in the Potomac River gorge near Weverton, Maryland.

Lithology

The Weverton is a complex lithologic unit in which Stose (1932, p. 43) recognized feldspathic sandstones, quartzites, and conglomerates in addition to some poorly exposed shaly beds. Good sections of this formation are nowhere exposed in the study area, thus making it difficult to reconstruct a stratigraphic sequence.

An attempt has been made to assemble a generalized section by complimenting the sections obtained in traversing East Big Flat Ridge and Piney Mountain along the power line adjacent to the Shippensburg-

Arendtsville road with good, local exposures of known stratigraphic positions. Although exposures along these traverses are sparse, abundant large blocks which have moved very little from true outcrop position, the absence of heavy plant cover, and the consistent attitude of the exposed strata aided in assembling a general lithologic sequence for this formation.

For purposes of description, the Weverton Quartzite can be divided into four lithologic intervals, comprising a basal, lower middle, upper middle and upper unit.

The basal unit of the Weverton is composed of coarse-grained, light-greenish-gray, phyllitic graywackes and quartzose graywackes with thin interbeds and lenses of dusky-blue color. Toward the top of this unit the bluish-gray color becomes more prominent and the rocks are, in part, cross-bedded. The beds and lenses of conglomerate containing dark phyllite and pink quartz pebbles about an inch in diameter are irregularly distributed through the lower and upper portions of the basal unit.

Lithologies at the base of the Weverton grade into the underlying conglomerate member of the Loudoun Formation by an increase in the regularity and thickness of the conglomerate beds as well as by an increase in the size and number of quartz, phyllite, and rhyolite pebbles. The writer (Fauth, 1962, p. 28) previously defined this contact at a lower stratigraphic level so as to include some of the coarse, polymictic conglomerate in the Weverton. Examination of the contact zone over a larger area indicates the contact as presently defined, excluding the more thick-bedded, coarse conglomerate beds, is more readily identified and more consistently mappable. In addition, this contact best approximates the contact of the Loudoun and Weverton recognized by other authors near the type area of the Weverton (Nickelsen, 1956, p. 248-49; Whitaker, 1955, p. 442-45, 459). The above relations are best observed near the base of Piney Mountain, along the power line north of the Shippensburg-Arendtsville Road.

At a locality along the south slope of the re-entrant on Piney Mountain about 1.2 miles west northwest of Strasbaugh School (Plate 1), the conglomerate member of the Loudoun Formation is absent. Here the Weverton is in an apparently conformable, sedimentary contact with the underlying Loudoun phyllites. Overlying the gray phyllites of the Loudoun, the basal Weverton consists of intercalated, thin-bedded, greenish-gray to bluish-gray quartzose phyllites and coarse-grained to pebbly, phyllitic quartzose graywacke of the same color. The graywacke commonly exhibits cross-bedding (Plate 2, Fig. 3). These lithologies grade upward into bluish-gray and light-gray, banded, and in part cross-bedded, phyllitic quartzose graywackes. The total thickness exposed is about 40 or 50 feet with the latter lithologies comprising about one-third.

Medium-grained, bluish and greenish-gray phyllitic quartzose graywacke and protoquartzite which becomes coarser and more quartzose upward, overlies the basal Weverton. Cross-bedding and irregular greenish-brown or yellowish zones occur in some of these beds. This is the lower middle unit of the Weverton.

Very coarse-grained to pebbly, greenish- or bluish-gray graywacke and protoquartzite with a few conglomeratic zones occurs in the lower half of the upper middle unit (Plate 2, Fig. 4). A graywacke conglomerate of rounded quartz pebbles about one inch in diameter held in a greenish phyllitic matrix, light-gray, very-coarse-grained quartzite, and banded, bluish- and greenish-gray graywacke with rectangular laths of weathered feldspar characterize the remainder of this zone.

The middle portions of the Weverton are not well exposed at any locality. Portions of this interval do outcrop, however, on the upper east slopes of Piney Mountain about two miles north of Strasbaugh School and also 9000 feet west southwest of Wenksville. Some exposures are also located along the crest of Rocky Mountain and at the north end of Wildcat Hill (Plate 1).

The rocks of the upper Weverton are mainly massive, coarse-grained, grayish-green protoquartzite and quartzite with thin interbeds of quartz pebble conglomerate and phyllitic sandstone occurring at wide intervals. Cross-bedding and fragments of weathered feldspar are found in some beds.

The uppermost Weverton is best exposed along the north side of the road in Dead Woman Hollow between the elevations of 1600-1640 feet. Exposures at this locality as well as elsewhere indicates a gradational contact with overlying unit, the Harpers Formation.

The sedimentary character of a representative group of samples from the Weverton is shown in Figure 6 and Table 2.

Thickness

The thickness of the Weverton in Pennsylvania has been estimated at various times by Stose as 750 feet (1932, p. 43) and 1250 feet (1906, p. 207; 1909, p. 4). The larger estimate may include rocks that more properly correlate with the Loudoun Formation (Stose, 1909, p. 4).

An earlier estimate by the writer (Fauth, 1962, p. 35) based on mapped outcrop width and attitude was 500-600 feet. Traverses along the power line adjacent to the Shippensburg-Arendtsville road (across the southeast slope of Piney Mountain and East Big Flat Ridge) suggests that the writer's original estimate is too small. The newly computed thickness, asuming a consistent high easterly dip as shown by the few exposures available, is 1550 feet for East Big Flat Ridge and 1630 feet for Piney Mountain (Plate 1). The upper contact was determined on float.

Recognizing the possibility of some small-scale folding and allowing for small errors in the location of the contacts, a thickness of 1200-1400 feet for the Weverton Quartzite is likely. On this basis thicknesses of the four lithologic intervals of the Weverton are: basal unit, 75-100 feet; lower middle, 250-300 feet; upper middle, 400-500 feet; upper, 400-500 feet.

Distribution

Five northeast-trending areas underlain by the Weverton Quartzite are shown on the geologic map (Plate 1).

The eastern belt is located along the summit and eastern slope of Piney Mountain, and extends from U.S. Route 30 to the northern border of the area.

East Big Flat Ridge and, in part, the summit and west slope of Wildcat Hill and the lowermost eastern flank of Yellow Ridge are underlain by Weverton Quartzite. This belt parallels the Conococheague Mountain Creek valley on the west for almost nine miles. Near the northern border of the area, Grave Ridge, to the east of Big Flat Ridge, is also underlain by this unit.

From the southern border of the area to U.S. Route 30, the Weverton occurs along the slopes and crestal portion of Rocky Mountain.

Big Pine Flat Ridge is developed on the Weverton. This belt extends from the northern border of the map to a point near the headwaters of Hosack Run in Dark Hollow. Stose (1929) mapped Weverton about one mile beyond Dark Hollow but the writer found no evidence to support this.

An irregular, elliptical area of Weverton has been mapped by the writer about midway between Methodist Hill and the westernmost ridges of South Mountain. This area, bounded on the south and north by Devil Alex Hollow and Furnace Run respectively, was mapped by Stose (1929) as Montalto Quartzite, but the evidence in float of the presence of Weverton is irrefutable.

Stose (1929) mapped a small body of Weverton in the valley occupied by Shirley Run. Because the author found no exposures or float of this formation to support Stose's mapping, the Weverton is not shown on the geologic map in Shirley Run.

Age and Correlation

No fossils have been found in the Weverton. Because of its conformable stratigraphic relationship to the other members of the Chilhowee Group, the upper portion of which contains Early Cambrian fossils, it is generally considered Early Cambrian.

Stose (1906, p. 204-05; 1932, p. 43-45;) and Stose and Stose (1946, p.

34-41) studied the stratigraphy of South Mountain at many localities in Maryland and Pennsylvania, and correlates the Weverton of the study area with that in the Potomac River gorge.

The writer has examined this unit at exposures in northern Virginia, West Virginia, Maryland, and near Waynesboro, Pennsylvania. Although the lithologic character of the unit changes across this region, the Weverton Quartzite in the Caledonia area maintains the same relative stratigraphic position as in the type area.

Harpers Formation

Name

The Harpers Formation was first described by Keith (1892, p. 363) and later named the Harpers Shale (Keith, 1893, p. 332) for extensive exposures along the Potomac River at Harpers Ferry, West Virginia.

Stose (1906, p. 206) recognized a thick section of quartzite in the Harpers in Franklin County, Pennsylvania. This interval was termed the Montalto Quartzite Member after Mont Alto, Pennsylvania where the quartzite underlies a series of prominent ridges.

Lithology

The best section of the Harpers Formation is along Route 16, east of Waynesboro, Pennsylvania. Cloos (1951, p. 34-36) has measured and described this section in detail. In general it consists of an upper interval of green, sandy, laminated phyllites and sandstones; a middle interval of interbedded sandstones, conglomerates, and phyllites; and a basal interval of interbedded bluish-gray or green phyllites and shales with some thin beds of sandstone.

Stose (1932, p. 45) described this formation as "gray hackly sandy shale, phyllite, or schist" which is largely replaced in northern Adams County by the Montalto Quartzite Member. This member is primarily a series of white, vitreous quartzites overlain by sandstones and quartzites containing numerous *Skolithos* tubes (Stose, 1932, p. 46). The Montalto Member thickens at the expense of the rest of the Harpers formation to the extent that the dark, shaly sandstones at the top and bottom of the formation are all that remains of the softer or more phyllitic beds of the Harpers type (Stose, 1932, p. 45).

In the Caledonia area, the Harpers Formation is divisible into three parts. The rocks above the Montalto Quartzite Member are, in this report, termed the upper Harpers; the Montalto Member is further subdivided into the lower Montalto and upper Montalto. Each of these lithologic units is a mappable body over the area covered by this investigation.

Since dark, silty or shaly sandstones typical of the basal portion of the Harpers Formation do not seem to exist below the Montalto Quartzite in the Caledonia area, the quartzite member immediately overlies the Weverton Quartzite.

The lower Montalto is rarely exposed. Rocks near the Weverton contact are exposed at two localities: west of the Piney Mountain road about one mile north of the Shippensburg-Arendtsville road, and near elevation 1680 just off the road in Dead Woman's Hollow. At these localities, grayish weathering, white, very-coarse- to coarse-grained protoquartzites, stained brownish along joint and bedding surfaces, occur near greenish protoquartzites and quartzites of the uppermost Weverton.

The remainder of this unit consists of white and grayish, medium-grained protoquartzites and quartzites which commonly are thin bedded and exhibit normal cross-bedding outlined by very thin, dark laminae. These rocks generally weather grayish orange to very pale orange, but a ¼ - 2 inch deep rind of grayish-red to brownish color is common.

Some of the protoquartzites contain small amounts of clayey material and altered feldspar. Minor amounts of hematite, occurring as small anhedral grains, are found in a few of these rocks. The hematite-bearing protoquartzites appear to represent a limited zone within the lower half of the lower Montalto, and are particularly common along the west flank of Graefenberg Hill (Plate 1).

White, medium- to coarse-grained protoquartzites and quartzites which weather gray but may show an orange or reddish stained periphery are found in the basal part of the upper Montalto. In some places, dark, silty or reddish (hematitic?) laminae parallel bedding.

Gray, vitreous quartzite, and fine- to medium-grained protoquartzite containing a few grains of weathered feldspar are exposed near the headwaters of Cold Spring Run. This lithology is generally overlain by a 15-25 foot thick interval of "bluish" protoquartzite or quartzite located near the middle of the upper Montalto. This "blue quartzite" lithology, best exposed near the foot of Green Ridge (Plate 1), consists of medium- to coarse-grained protoquartzite and quartzite which contain small specks of hematite or hematite-rich lenses or bands up to one-half inch thick. On weathering, the rock shows a grayish to grayish-blue, or less commonly, a grayish-red color.

The "blue-quartzite" lithology is traceable throughout most of the area studied. It has been used extensively by the writer to outline structural patterns and as an aid in locating the upper contact of the Montalto Quartzite Member. The unit is shown on the geologic map (Plate 1) and Figure 18.

Stratigraphically above the "blue quartzite" are beds of medium- to coarse-grained, yellowish-orange weathering, grayish quartzite and protoquartzite. Laths of weathered feldspar occur in some of these beds which grade upward into finer-grained, brownish-yellow to olive-gray protoquartzite and quartzite.

On the west flank of South Mountain near Stony Knob, the "blue quartzite" interval is absent, and the uppermost Montalto is composed of medium-grained, very-light-gray to light-gray quartzite with irregular zones which weather pale red or pinkish gray.

Limonite-stained, greenish-gray, very-fine-grained graywacke and quartzose graywacke, weathering to shades of olive gray and orange brown characterize the upper Harpers.

Figure 6 and Table 2 summarize the sedimentary characteristics of this formation.

Thickness

Stose (1932, p. 46) estimates a thickness of 3000 feet for the Harpers Formation in Adams County and 2750 feet for an area just west of the county. In the same report, about 75 per cent of the thickness is referred to as quartzite; presumably this represents the Montalto Member which is then assigned a thickness of 2050 feet.

Cloos (1951, p. 33-36) has measured 3100 feet of Harpers Formation east of Waynesboro, Pennsylvania, but suggests that exaggeration of thickness due to folding may be considerable.

The thickness of the Harpers Formation and its subdivisions is estimated on the basis of the outcrop pattern and attitude of these rocks in the area studied. Outcrop pattern along the western slope of Piney Mountain and in the vicinity of Green Ridge suggests a thickness of 900-1100 feet for the lower Montalto, and 700-800 feet for the upper Montalto. The total thickness for Montalto Member is then 1600-1900 feet. On the basis of a pace and compass traverse across the lower slopes of Green Ridge, the "blue quartzite" zone is placed approximately 300-400 feet below the contact with the upper Harpers.

A thickness of 300-500 feet for the upper Harpers is suggested by exposures northwest of Wigwam Hill and Quarry Hill (Plate 1). Thus, the Harpers Formation has an approximate thickness of 1900-2400 feet in the Caledonia Park area.

Distribution

From Graefenburg Hill on the south, the Harpers Formation extends northeastward along the western slope of Piney Mountain. The occurrence of the upper Harpers east of Mountain Creek near the northern border of the area is limited by faulting and only the Montalto Member is present. The "blue quartzite" is found in the southern half of the belt and also at the northern end. Locally where it is absent, faulting is suspected.

The Harpers is represented by the Montalto Member in a downdropped block not recognized by Stose (1929) about one mile east of Caledonia Park at the southern tip of Piney Mountain (Plate 1).

Ludwigs Hill and Kettle Springs Mountain, located just west-southwest of Caledonia Park, are the northern extension of Montalto Mountain, the type area of this member of the Harpers Formation. A complete section of the formation is found along these ridges.

A narrow band of Montalto, primarily the lower unit, occurs along the lower slopes west of Conococheague Creek. This band extends from the vicinity of the Miltonburger Road southward to the foot of Yellow Ridge where it joins a wider belt of Harpers which underlies much of the western slope and summit of East Big Flat Ridge and eastern flank of Yellow Ridge. The merged belts then comprise all of Chinquapin Hill and most of the knobs to the southwest.

Valleys developed on the upper Harpers outline a broad, U-shaped belt which encloses all of the upland zone in the western half of South Mountain (Plate 1). Except for sections around Big Flat Ridge and in the vicinity of Furnace Run, the upland area is underlain by the Montalto Member. Only locally is the upper Harpers missing along the periphery, probably as a result of faulting.

The "blue quartzite" which occurs along the eastern and southern margins, is conspicuously missing on the western flank except for the northwesternmost corner of the map area. Whether the absence of this unit is structural or stratigraphic is uncertain, but this represents the longest single belt along which this unit is absent. The "blue quartzite" unit was inadvertantly omitted from the Geologic map (Plate 1) west of Chinquapin Hill and Yellow Ridge although it was mapped there. The trace of the "blue quartzite" in this area is shown in Figure 18.

Age and Correlation

Fossils other than *Skolithos* tubes have not been found in the Harpers. It is generally assigned an Early Cambrian age because of its conformable stratigraphic relation to the overlying Antietam which contains Early Cambrian fossils.

Skolithos tubes, restricted to the Montalto Quartzite Member, occur in varying amounts throughout much of this unit.

The writer has traced the Harpers, including the Montalto Quartzite Member, into the map area from the type area of the Montalto. Sections of the Harpers formation have also been examined near Waynesboro, Pennsylvania and Harpers Ferry, West Virginia. The writer is in agreement with Stose (1932, p. 46-47) and Stose and Stose (1946, p. 40-41) that the Harpers Formation in the map area has the same relative stratigraphic position as those rocks termed Harpers in the type area.

Antietam Quartzite

Name

The Antietam Quartzite was recognized and named by Keith (1892, p. 365). Cloos (1951, p. 39) states that the type locality is along "the tributaries of Antietam Creek, Washington County, Maryland".

Lithology

Stose (1932, p. 47) describes the Antietam of the South Mountain region as a clean, coarse-grained, quartzose sandstone divisible into two lithologic subdivisions. The lower division is a resistant, bluish to pink quartzite, and the upper member consists of white to pinkish sandstone containing numerous *Skolithos* tubes.

The Antietam is normally well-exposed in the study area, but nowhere is there a complete section available, and neither top nor bottom are exposed. However, in adjacent areas these contacts are exposed. The contact with the underlying Harpers as exposed along Route 16 east of Waynesboro is gradational, and the occurrence of fine siliceous "schist" near the base of the younger Tomstown Dolomite in an old iron pit near Mont Alto (Stose, 1932, p. 49) suggests that the upper boundary is also transitional.

Exposures located along the hogback ridge northwest of Caledonia Park, at the south end of Stony Knob, and in the two quarries east of English Valley are the best localities for determining the lithologic character of this unit.

At the northern end of Quarry Gap Hill about 100-120 feet of Antietam is exposed in a series of outcrops, each about 10-15 feet thick. The lower 60-70 feet stratigraphically occurs near the ridge crest. The rock is basically a medium-grained, light-gray protoquartzite which weathers gray to grayish orange. The weathered surface is friable and usually contains circular depressions ¼ - ¾ inch deep. *Skolithos* occurs in some beds.

The next highest zone, 15-20 feet thick is composed of slightly friable, medium-grained protoquartzite containing minor amounts of clay. The weathered rock is commonly pitted and has a very-pale-orange color.

About 30 feet of fine to medium-grained, pale-orange weathering protoquartzite and quartzite complete the section. These beds tend to be thin (2-8 inches), slightly friable, and contain fine laminae of hematite. *Skolithos* tubes occur locally.

Along the stream at the southwestern tip of Stony Knob approximately 160-180 feet of upper Antietam is exposed. The section is composed almost wholly of white to yellowish-gray, medium-grained protoquartzite and quartzite. These rocks are slightly friable and usually contain minor amounts of clay. The clay, most noticeable on the weathered surface, occurs as irregular thin zones or evenly disseminated throughout the rock. On weathering the rocks develop a gray or pale-orange color. Limonite stains are located along fractures. *Skolithos* tubes, some 10 inches long, are found throughout much of this section (Plate 2, Fig. 5).

On the second highest level of the Mountain Cydonia Quarry, a 75-90 foot section of white to very-light-gray to grayish-orange-pink (10 R 8/2), medium- to coarse-grained quartzite is exposed. Most of the beds are about 6-12 inches thick and contain *Skolithos* tubes normal to bedding. A friable character and yellowish-orange color are typical of the weathered rock.

Beneath these beds is a 25-30 foot section of very-light-gray, medium- to coarse-grained, friable protoquartzite and quartzite containing specks of clay. At intervals of seven to nine feet, one to two inch beds of bluish-gray clay occur.

The southern end of the Caledonia Quarry exposes *Skolithos*-bearing, white to light-gray, medium-grained quartzite interbedded with white, coarse-grained, friable protoquartzite containing minor amounts of clay. The total section is about 30-35 feet thick, and each lithologic sub-unit has a thickness of five to eight feet.

Quartzites of light-gray color occur in parts of the Antietam. Generally this rock is medium-grained, but coarser varieties may occur. Specks of hematite or limonite are commonly observed and give the rock a grayish-orange-pink (10 R 8/2) to grayish-red color (5 R 4/2) on exposed surfaces.

Locally, as at the mouth of Dark Hollow, yellowish-brown to light-brownish-gray, fine- to medium-grained quartzose graywacke and protoquartzite crops out. The bedding is 4-15 inches thick, with the thinner beds characterized by irregular or wavy bedding surfaces.

A quartz-pebble conglomerate, located near the top of the formation, is developed in places. An exposure at the base of Quarry Hill consists of well-rounded, white to light-gray quartz pebbles, 0.2 - 0.8 inches in diameter, in a matrix of coarse-grained, grayish, orange-stained quartz. A quartz pebble conglomerate is also locally developed at this same stratigraphic level in the Antietam in northern Virginia.

Figure 6 and Table 2 summarize the sedimentary characteristics of the Antietam.

Thickness

Stose (1932, p. 48-49) assigned a thickness of 800 feet to the Antietam in Adams County. Based on mapped outcrop width and dip at a number of localities, especially at Stony Knob and Quarry Gap Hill, the author estimates a thickness of 700-900 feet for the area studied.

Distribution

Except for a limited area of exposure in the Graves Ridge-Mountain Creek area, the Antietam is restricted to the area west of Route 233.

This formation underlies the discontinuous series of ridges which encircle the Big Flat Ridge-Pine Flat Ridge upland, the westernmost ridge of South Mountain southwest of Mount Union, and Little Mountain.

Sandstones and quartzites located on the east flank of Chinaquapin Hill were identified by Stose (1929) as Antietam. However, the position of the "blue quartzite" unit of the Montalto (Figure 17) and the structural pattern it outlines demonstrates that these rocks belong to the lithologically similar Montalto Quartzite.

On the basis of the position of the "blue quartzite" unit near Strasbaugh Hill quartzites mapped to the west of this hill must be Antietam and not Montalto as shown by Stose (1929).

A similar misidentification resulted in the mapping of a larger belt of Antietam along the eastern flank of Grave Ridge. Structural evidence and lithology again demonstrate that the Montalto underlies most of this zone.

Age and Correlation

Fossils other than *Skolithos linearis* have not been found in the Antietam Quartzite of the study area. Elsewhere in Pennsylvania and Maryland, *Camerella minor, Hyolithes communis, Obollela* sp., *Olenellus* sp. have also been recognized in these rocks (Walcott, 1896, p. 25; Bassler, 1919, p. 59; Resser, 1938, p. 5; Amsden, 1951, p. 102). Thus, an Early Cambrian age is assigned to this formation.

Tomstown Dolomite

Name

This formation was named by Stose (1906, p. 208; 1909, p. 5) for scattered exposures near the village of Tomstown, Franklin County, Pennsylvania.

Lithology

The Tomstown Dolomite is poorly exposed throughout the South Mountain belt of Pennsylvania. The absence of Tomstown exposures is an effect of its topographic and stratigraphic relationship to the Antietam Quartzite, whose debris forms an extensive colluvial-alluvial apron west of the Blue Ridge front, and of the soluble character of the Tomstown lithologies.

From the type area, Stose (1929, p. 7) reports shales interbedded with dolomite from the basal portion of this formation, thin bedded, dark-blue limestone near the middle, and massive blue magnesian limestone containing some black chert at the top.

In the area studied by Farlekas (1961,p. 38-39) at the southern end of South Mountain in Pennsylvania, thin-bedded, dark-bluish-gray dolomitic limestone with paper-thin argillaceous partings is the principal lithology exposed.

Drilling near Frederick, Maryland penetrated about 180 feet of Tomstown before entering the Antietam. The lower 40 feet is a white to light-gray, medium- to fine-grained, massive to thin-bedded dolomite with minor sericitic partings, and the upper 140 feet, immediately beneath a fault, consists of gray, thin-bedded dolomite limestone with frequent sericite and carbonaceous partings (Hoy and Schumacher, 1956, p. 1525).

Nowhere in the area studied by the writer does the Tomstown outcrop or occur in float. Its presence has been inferred on the basis of the stratigraphy, and the topographic and photographic expression of the formation. The location of iron-ore pits has also been utilized as they tend to occur near the base of the formation (Stose, 1932, p. 139-141).

Thickness

In Franklin County, Stose (1932, p. 49) assigned an estimated thickness of 1000 feet to the Tomstown. Cloos (1946, p. 40-41) cautioned that the thickness attributed to the Tomstown Dolomite and the overlying limestones can hardly be reliable in view of the position of these formations to the area of intense folding and oolite deformation.

Distribution

The Tomstown is inferred to occur in three separate zones within the study area. The largest zone is a belt extending along the entire length of the western flank of South Mountain. The presence of abundant iron-ore pits, the presence of features interpreted as solution features on aerial photographs, and the topographic low between the westward dipping Antietam and an irregular line of low ridges formed by the overlying Waynesboro Formation, notably in the southern part of the area, are the evidence used to extend this formation northward along the flank of the mountains from its type locality (Plate 1). This zone is appreciably narrowed west of Little Mountain by faulting.

The Tomstown is mapped in the central part of the area, east of the series of ridges formed by the Antietam (Quarry Hill to Wolf Hill), and in the valley of Mountain Creek, near the northern margin of the map area (Plate 1) on the basis of the stratigraphy, solution features observed in the field and on aerial photographs, and the occurrence of iron-ore pits. The extent of the Tomstown in the Mountain Creek valley is suggested to be less than previously shown (Stose, 1929; Fauth, 1962) because the Montalto Quartzite Member, in part, was misidentified as Antietam Quartzite (Plate 1).

The area of Tomstown Dolomite mapped by Fauth (1962) just west of Caledonia Park is probably in error. An excavation about 0.7 miles west of Caledonia and just north of U.S. 30 uncovered vertical beds of Montalto striking directly across this zone. Rather than representing a down-dropped block of Tomstown it is now assumed that this area is underlain by the normal Chilhowee section.

Age and Correlation

The Tomstown is sparingly fossiliferous. In the upper limestones of the unit near Roadside, Franklin County, Walcott is reported to have found specimens of *Salterella* sp. undet., *Kutorginia,* and fragments of *Olenellus* (Stose, 1906, p. 699). This evidence suggests an Early Cambrian age for this unit. The Tomstown Dolomite is considered an equivalent to the Shady Dolomite of northeastern Tennessee (Table 2).

Waynesboro Formation

Name

The name Waynesboro Formation is applied to this unit for conspicuous outcrop in the ridge just north of Waynesboro, Franklin County, Pennsylvania (Stose, 1906, p. 209).

Lithology

The Waynesboro is a series of shales, sandstones, and limestones which overlie the Tomstown Formation. As reported by Stose (1908, p. 700), the base of the unit consists of sandy gray limestone that weathers to a slabby, porous sandstone, large masses of rugose chert, and white vein quartz. Dark-blue to white limestones and dolomites occur in the middle of the formation, grading into mottled, slabby sandstones and purple siliceous shales toward the top.

Cloos (1946, p. 42-43) describes a partial section of this formation from Chewsville, Maryland which, in general, conforms with the description given by Stose for the Pennsylvania section.

Exposures of the formation are limited to the low hill west of Little Mountain, and along the edges of the road from Pond Bank to Duffield, beginning about 0.6 mile west of Pond Bank (Plate 1). Elsewhere this unit is concealed by Antietam debris.

West of Littlt Mountain, a questionable outcrop of medium-gray limestone is overlain by yellowish-orange weathering, laminated sandstone and silty shale, and a zone of purplish- and olive-colored shales. A similar sequence occurs along the road west of Pond Bank. Starting near the creek and continuing for about 1500 feet west, the following lithologies occur in float or in small exposures along the sides of the road; medium-bedded, laminated, gray, crypto-crystalline limestone; friable, fine-grained sandstone and greenish-gray silty shale; olive-colered, yellow-orange weathering shales; thin-bedded, medium-gray limestone; and interbedded olive and purple shales.

Thickness

"The thickness of the formation, computed from the width of outcrop and dips, and allowing for minor folding, is 1250 feet" (Stose, 1908, p. 700). It is impossible to determine the thickness of the Waynesboro in the Caledonia Park area because, where exposed, the lower part of the formation is absent owing to faulting (Plate 1). The estimated thickness of the incomplete Waynesboro section west of Little Mountain, based on mapped outcrop width and dip, is 900-1100 feet.

Distribution

The Waynesboro Formation occurs in a belt along the eastern edge of the Cumberland Valley. Its relatively resistant character causes it to form low knobs and ridges parallel to the general trend of the mountain front, which usually provides a means of tracing the formation northward even in areas of poor exposure.

The lower part of this formation is not exposed in the area studied. Generally it is covered by colluvial material or, as in the vicinity of Little Mountain, is faulted out against the underlying rocks (Plate 1). The purple shales of the upper Waynesboro are the only readily traceable lithology. However, north of Little Mountain, even they are covered. Thus, only a very general knowledge of the distribution of the formation is known north of this locality.

Age and Correlation

Stose (1908, p. 700) reports identifiable remains of the brachiopod *Lingullela* from the sandy shale near the top of the formation near Waynesboro, suggesting a Middle Cambrian age. No fossils have been reported from this formation in Maryland, however, beds of equivalent stratigraphic position in Virginia (Rome Formation) contain the *Olenellus* trilobite fauna (Amsden, in Cloos, 1946, p. 102).

Elbrook Formation

Name

The Elbrook Formation was named by Stose (1906, p. 209) for exposures near the town of Elbrook, Franklin County, Pennsylvania.

Lithology

The Elbrook includes massive magnesian limestone with prominent red and green shales in the middle, and beds of sandy limestone in the upper part of the unit (Stose, 1906, p. 209).

Wilson (1952, p. 304) describes this formation as a sequence of yellow weathering dolomitic and calcareous shales which alternate with limestone in no apparent, regular lithic succession.

Because the Elbrook commonly occurs near the western flank of South Mountain, its basal part is normally covered by an extensive apron of colluvial material. The gross lithology of the lower portion of this formation can only be studied along the Pond Bank-Duffield Road near the southwestern corner of the area. Owing to the finely laminated and shaly character of the lower Elbrook, good exposures are rare and much of the data are only obtainable from float or small, two-dimensional outcrops along the roadside.

The lowermost Elbrook is similar lithologically to the upper Waynesboro Formation. The contact is drawn above the last purplish shale of the Waynesboro. On that basis the lower Elbrook is primarily an interval of medium-grayish-green shales or calcareous shales which weather to yellow-orange or mustard-colored plates, and subordinate, thin-bedded, fine-grained, gray limestone. This lithologic combination appears to predominate throughout much of the lower half of the unit.

Yellow-orange weathering, greenish and purplish shales form a moderately thick interval toward the middle of the Elbrook. In many places, thin, black laminations in the greenish shales serve to outline bedding. These beds crop out along Route 997 about midway between East Fayetteville and Duffield.

Scattered exposures and float of the Elbrook west of Duffield consist mainly of thin-bedded, light-gray limestone and greenish-gray calcareous shales which weather yellow brown. Thin silty laminae which weather yellowish brown are commonly developed in the limestone. Calcareous sandstone which become friable and yellow brown on weathering, and fine-grained sandstone occur in thin beds scattered among the laminated limestone and shale.

Numerous small outcrops of the Elbrook occur in fields west of East Fayetteville and south of Route 30. Judging from the position of these exposures relative to the Conococheague-Elbrook contact mapped by Stose (1909), these beds lie in the upper third of the Elbrook. Light-gray weathering, medium- to dark-gray, dense limestone with paper-thin laminae of calcareous shale or very thin silty beds are predominant. The laminations usually weather yellowish gray or very pale orange.

Good partial sections of this formation are found in an abandoned quarry on Conococheague Island and along the creek at this same locality. The quarry exposes thick-bedded, light-gray weathering, medium-gray, cryptocrystalline to fine-grained limestone, and beds of buff weathering, light-gray dolomite. A section about 200 feet thick is exposed along the south side of the creek. The sequence consists of thick-bedded, grayish-yellow weathering, impure limestone containing ¼ - 1 inch intervals of clayey limestone; fine-grained sandy limestone; orange-brown weathering, laminated to mottled, light-gray, cryptocrystalline limestone; and thin-bedded, grayish-yellow, calcareous shale.

Bedding-cleavage relations suggest the rocks in the quarry are older than those at the creek, and may represent the medial limestone noted by Stose (1909, p. 5-6) which are exposed along the strike in the vicinity of Woodstock. Westward along the creek, the rocks become younger.

Rocks comprising the upper Elbrook are exposed in a series of small outcrops east northeast of Scotland, just west of Route 696. These beds occur within 400-1000 feet of the basal units of the overlying Conococheague Limestone. Since most of the dips are between 15 and 40 degrees, the stratigraphic separation is probably between 100 and 400 feet.

In this zone, the lithic succession is complex because many lithologic types occur in beds only a few feet thick and there is no recognizable order to the lithic succession.

The most prominent lithologies present in the upper Elbrook are: thin, yellowish-gray calcareous shale; medium-light-gray weathering, dense limestone; yellowish-gray weathering, pinkish-gray, sandy limestone; and grayish, fine-grained, laminated limestone. Grayish, oolitic limestone and light-olive-gray weathering, edgewise conglomerate composed of limestone plates about one inch long in a dark-gray, finely crystalline matrix also occur.

Although the contact between the Elbrook and Conococheague is not exposed, the similarity in lithology of the beds near the contact suggests a conformable and transitional boundary. Wilson (1952, p. 304) states "Conococheague-type limestone has been noted from the very top of the Elbrook to as low as 800 feet below the base of the Conococheague".

Thickness

The thickness of this formation is not easily estimated for the area because of poor exposure and lack of structural control. Independent calculations of the thickness made on the limbs of an asymmetrical syncline near Quincy, Franklin County, Pennsylvania suggest a value of about 3000 feet (Stose, 1909, p. 5).

Distribution

The Elbrook is confined to the western quarter of the area where it occupies a broad band approximately parallel to the mountain front at a distance of 1 - 1½ miles.

Age and Correlation

Other than rare stromatolites, the writer has not recognized any fossils in this unit. Fragments of trilobites reportedly obtained from some of the purer limestones at a quarry near Waynesboro suggest a Middle Cambrian age for this formation (Stose, 1909, p. 5).

Conococheague Formation

Name

The rocks of this unit were originally described as part of the Knox Formation (Stose, 1906, p. 209-210), but the lower portion of this formation was later distinguished as a separate formation by Stose (1908, p. 703) and named Conococheague for exposures in the town of Scotland along the Conococheague Creek.

In the Greencastle and Waynesboro 7½-minute quadrangles immediately southwest of the study area, the Conococheague is raised to group status (S. Root, 1968) and two formations, a lower Zullinger and an upper Shadygrove are mapped. The Shadygrove corresponds to the upper part of the Conococheague as mapped here and the Zullinger Formation represents the remaining portion.

Lithology

The type section of the Conococheague is adjacent to the northwest corner of the study area. A section of the Conococheague was measured by Stose (1908, p. 702) at this locality.

Wilson (1952, p. 306-307) notes that the section at Scotland is incomplete; about 500 feet of the lower Conococheague beds present east of the village of Scotland do not occur in the type section. The Waynescastle Dairy section, described by Wilson (1952, p. 305-307; 313-315), is the most complete section of this formation in Pennsylvania.

The basal Conococheague beds tend to form a low but relatively persistent ridge northeast of Scotland. The dominant lithology is a yellow-orange (10 YR 8/6) weathering, coarse- to very-coarse-grained calcareous sandstone to arenaceous limestone. In outcrop this rock is deeply weathered and crumbles readily into loose sand. Associated with this rock are grayish, very-fine-grained limestones, and dark-gray, laminated, cryptocrystalline limestone. The laminae are dolomitic bands ½-2 inches thick which weather grayish orange or yellow gray. On the weathered surface, the dolomite stands in positive relief.

This lithologic interval has been named the Big Spring Station Member by Wilson (1952, p. 307). Although a recognizable lithology, this portion of the Conococheague is commonly covered by the distal end of an extensive gravel deposit (Plate 1), and therefore has not been mapped separately.

The contact between the Elbrook and the Conococheague is not exposed in the area, but Wilson (1952, p. 308) states that it is transitional.

Good exposures of the lower middle part of this formation occur in a railroad cut north of Scotland. The rock is mainly a light-gray weathering, medium- to dark-gray, banded, cryptocrystalline limestone. The bands, ¼ - 3 inches thick, are composed of grayish, silty, dolomitic limestone which weathers yellowish brown to grayish brown. Occurring every few inches, the dolomite zones give the rock a banded or "ribbed" character since they weather into positive relief (Plate 2, Fig. 6).

Flat-pebble conglomerate and finely laminated gray limestone are subordinate and usually intercalated with the "ribbed" limestone. These beds become more abundant in the upper middle Conococheague.

The upper midde part of the Conococheague Formation consists of the following major lithologic units:

(1) Dark-gray oolitic limestone which weathers light gray; grades vertically into intraformational limestone conglomerate with an oolitic or calcarenitic matrix; some thin intercalations of silty limestone.

(2) A complex association of calcarenitic, algal, and oolitic limestones with very subordinate thin beds of limestone pebble conglomerate; the limestones are mainly dark gray but weather light gray.

(3) Dark-gray, cryptocrystalline, light-gray weathering limestone with ¼ - ½ inch thick bands of yellowish-gray weathering, gray silty dolomite occurring at intervals of a few inches; very fine laminae of silt outline complex current bedding.

(4) Finely interlaminated gray, cryptocrystalline limestone and buff-weathering, very fine silty dolomite.

(5) Finely crystalline, buff-weathering dolomite; thin zones of sandy dolomite and thin dolomitic sandstone beds are subordinate.

These lithologies comprise cycles 5-15 feet thick which occur, with little local modification, throughout the outcrop belt of this portion of the Conococheague Formation in the study area. Similar cyclicity in the Conococheague is recognized and described in detail by Root (1964) for an extensive area of the Cumberland Valley south of the Scotland 7½-minute quadrangle.

The uppermost portion of this formation, the Shadygrove Formation of Root (1968), is characterized by cryptocrystalline, pink limestone and nodular black cherts. These beds immediately overlie an interval of poorly exposed cherts, flat-pebble conglomerate, and silty or sandy limestone. These lower cherts are conspicuous in float as they weather to a grayish-orange, rough, and porous mass.

The upper Conococheague grades into the overlying Stonehenge Formation. However, the pinkish limestone and black cherts make the contact zone readily distinguishable in the field.

Thickness

The thickness assigned to the Conococheague Formation from exposures at the type locality is 1600 feet (Stose, 1908, p. 702). Wilson (1952, p. 306-307) suggests the figure is too small since about 500 feet of basal Conococheague is absent from the type locality. The corrected thickness value is about 2100 feet.

Wilson (1952, p. 307) has measured partial thicknesses of 1890 and 1900 feet for this formation at the Waynescastle Dairy and the Big Spring Station sections respectively.

On the basis of outcrop width and attitude, an estimate of at least 2000 feet is justified.

Distribution

The Conococheague Formation occurs as a band about one mile wide in the northwest corner of the map area.

Age and Correlation

Fossils identified by Stose (1908, p. 703) from the base of the formation near Scotland include the trilobites *Dikelocephalus hartii* and *Billingsella* cf. *B. desmopleura;* both prove a Late Cambrian age. The fauna of the Conococheague, as defined by Stose, indicate this unit almost spans the Late Cambrian Epoch (Wilson, 1952, p. 315-316).

Sando (1958, p. 839-841) redefined the base of the overlying Stonehenge Formation to include the thin-bedded limestones of the upper Conococheague as recognized by Stose, because these rocks contain a small but diagnostic Lower Ordovician trilobite fauna. Both *Symphysurina* and *Clelandia* occur in this zone and characterize a basal zone of the Lower Ordovician (Sando, 1958, p. 840).

The author was unable to clearly and consistently distinguish the Conococheague-Stonehenge contact as defined by Sando, perhaps owing to the limited extent of contact zone in the study area (Plate 1). Consequently the author reverts to the original definition of the top of the Conococheague as described and mapped by Stose (1908, p. 6; 1909), and later utilized by Wilson (1952, p. 308-311). It is drawn about 120-150 feet above a black chert and fine-grained, pinkish-gray limestone interval which seems to represent a continuous and distinctive lithologic horizon. Thus, the Conococheague Formation as defined by Stose and mapped by the author, includes most of the Upper Cambrian and the basal portion of the Lower Ordovician.

ORDOVICIAN ROCKS

Stonehenge Formation

Name

The name Stonehenge was introduced by Stose (1908, p. 704) for the basal member of the entire Lower Ordovician section, the Beekmantown Limestone, but was later elevated to formational rank by Sando (1958, p. 879).

The name was taken from the village of Stonehenge, now included in Stoufferstown, immediately east of Chambersburg, Franklin County, Pennsylvania.

Lithology

Stose (1908, p. 6) described dark- to light-gray massive limestone and upper, fine-grained gray limestone with silty and sandy laminae from the base of a section of Beekmantown Limestone east of Chambersburg, and assigned these rocks to the Stonehenge. The fine-grained, massive limestone marks the contact with the underlying Conococheague Formation which is characterized by sandy, conglomeratic, thin-bedded limestone (Stose, 1908, p. 6). Sando moved the base of the Stonehenge as defined by Stose stratigraphically downward to include about 200 feet of strata formerly considered part of the Conococheague because "this arrangement is a better expression of temporal unity and is no less useful in geologic mapping" (Sando, 1958, p. 839). The name Stoufferstown Member was assigned this newly defined part of the Stonehenge Formation by Sando (1958, p. 839).

The writer could not consistently identify and map the Stoufferstown Member in the study area, consequently the base of the Stonehenge in this report is the same as that defined and mapped by Stose (1908, p. 6; 1909). Root (1968) was able to map the Stoufferstown over a wide area in the Greencastle and Waynesboro 7½ -minute quadrangles immediately to the south. The Stoufferstown is thus raised to a formation in that area.

Exposures near the base of the Stonehenge Formation are mainly light-gray, massive, cryptocrystalline limestones with gray, finely laminated limestone and calcarenite above. The remainder of the formation is characterized by light- to medium-gray weathering, medium- to dark-gray edgewise conglomerate, and interbedded gray, thin-bedded, very-fine-grained limestone and dark-gray to black calcareous shales. The limestone plates in the conglomerate are usually one-half to three or four inches long and one-quarter inch thick. The matrix is a calcarenite containing silt-size carbonate detritus and fossil debris.

Near the top of this unit, yellowish-gray weathering dolomitic seams and bands in the grayish limestones weather in positive relief, and thus these rocks bear a strong resemblance to rocks of the Conococheague Formation.

Thickness

The total thickness of the Stonehenge in the type area, about five to six miles southwest of exposures in this study area, is about 1000 feet (Sando, 1958, p. 840). Since this figure includes about 200-225 feet of Stoufferstown, the thickness of the unit defined by Stose and used by this writer is about 800 feet.

An estimation of the thickness of the Stonehenge Formation based on outcrop width and dip of the rocks in the study area is 675-750 feet — a figure which corresponds closely to that of the upper member of the Stonehenge as defined by Sando in the type area.

Distribution

A northeast-trending band less than 1000 feet wide near the northwestern corner of the study area is the only occurrence of the Stonehenge Formation.

Age and Correlation

Few identifiable fossils occur in the Stonehenge Formation. Sando (1958, p. 841) reports *Bellefontia collieana, Finkelnburgia bridgei, Finkelnburgia stonehengensis*, and lytospirid gastropods as the most common forms in the upper member, with *Tentaculites lowdoni* occurring near the top. The Stoufferstown member yields *Symphysurina* cf. S. *woosteri* and *Clelandia* sp., both of which characterize a zone near the base of the Lower Ordovician (Sando, 1958, p. 841). Hence, the Stonehenge Formation as mapped by the writer is of Early Ordovician age.

Rockdale Run Formation

Name

The Rockdale Run Formation was named for exposures along Rockdale Run near Hicksville, Washington County, Maryland by Sando (1957, p. 21). This unit represents the upper portion of the Beekmantown Limestone as described by Stose (1908, p. 6-7).

Lithology

Because of the limited areal extent and exposure of this formation in the study area, the description of the Rockdale Run is mainly from Sando (1958, p. 839-841, 845-850).

The base of the unit is defined on the occurrence of a laminated dolomitic limestone (Sando, 1958, p. 839). A bed of similar lithology, one foot thick, occurs just inside the western boundary of the map, 1.8 miles north of Scotland, and is the basis for recognizing this formation in the map area.

Sando (1958, p. 841) describes the remainder of the formation above the dolomitic rock as predominantly algal and mechanical limestones in the lower two-thirds, and dolomite and dolomitic limestone in the upper third.

In the area mapped, scattered outcrops above the basal beds are mainly light-gray weathering, dense, dark-gray limestones. These beds contain irregularly distributed, very thin clayey laminae which stand in positive relief on the weathered surface. Associated with this lithology are subordinate grayish-yellow weathering, fine-grained limestone, and grayish flat-pebble conglomerate. The upper dolomitic limestones do not occur within the study area.

Thickness

Sando (1958, p. 841) has measured an incomplete section 2547 feet thick near Stoufferstown. The Rockdale Run equivalent in a section described by Stose (1908, p. 6) near Chambersburg has a thickness of about 1680 feet. In the area mapped where the top of the formation is not exposed, a minimum thickness figure of about 1800 feet is valid on the basis of outcrop width and attitude of the rocks.

Distribution

The Rockdale Run Formation is restricted to the northwestern corner of the map area.

Age and Correlation

No identifiable fossils were recognized by the writer in the rocks exposed in the study area. However Sando (1958, p. 841-842) has collected and identified numerous fossils near Chambersburg including *Isoteloides* cf. *I. flexus, Tentaculites lowdoni,* and *Finkelnburgia* sp. The fauna suggests an Early Ordovician age for this formation, although the uppermost and unfossiliferous portion may represent the base of the Middle Ordovician (Sando, 1958, p. 845).

CRETACEOUS (?) TERTIARY - QUATERNARY DEPOSITS

Terrace Gravels (Alluvial-Colluvial Deposits)

Character and Distribution

Extensive terrace gravel and colluvial deposits are developed in the South Mountain area. The most notable deposit of this type occurs at the base of the westernmost ridges, usually between the 700 and 850 foot contours. The deposits extend westward for about 1-1½ miles, although some extend as much as 3 miles locally.

An examination of the shape and distribution of the deposits on the topographic maps and aerial photographs suggests that they mainly represent a series of alluvial fans. That the fans have been partially dissected is demonstrated by the 20-60 feet deep valleys occupied by the present streams which flow across these deposits.

The terrace gravels and alluvial-colluvial deposits are composed of materials derived from the rocks which form the adjacent ridges. The main components are subrounded to rounded cobbles, pebbles and sand grains made up of pinkish, yellowish, and grayish sandstone and quartzite. Grain size in general increases toward the mountain front.

With increasing depth these deposits tend toward a sandy clay, and the grayish to yellowish-brown color of the upper portion becomes distinctly reddish brown. The bulk of the clay in this zone is believed to be derived from the weathering of the underlying carbonates.

Local, thin iron-rich zones and thin beds of lignite material of unknown areal extent occur. The lignite is reported from the vicinity of Pond Bank (Frazer, 1875, p. 250; Pierce, 1965).

Thickness and Age

The thickness of the terrace gravels and alluvial-colluvial deposits decreases to zero outward from South Mountain. The maximum thickness attained by these deposits is not known. However, a 70-foot shaft developed by an iron mining company near Pond Bank (Lesley, 1865, p. 481) did not strike bedrock. Information gained from other prospect pits and shafts located in this material along the mountain front all indicates a general maximum thickness in excess of 40 feet.

Additionally, both Conococheague Creek just west of Caledonia Park, which is incised at least 60 feet, and the south fork of Phillamen Run northwest of Eagle Rock, which has cut down at least 80 feet, have not penetrated these deposits. K. Pierce (1965, p. 153) suggests a thickness of at least 170 feet or as much as 335 feet for the Pond Bank locality based on information in the reports of E. V. d'Invilliers (1887) and Lesley (1865).

The age of these deposits is unknown. Stose (1909, p. 13) suggests a Tertiary to Recent age on the basis of physiographic evidence. Recently fossil pollen and spores from the surficial deposits at Pond Bank have been identified by R. H. Tschudy as Late Cretaceous in age (Pierce, 1965, p. 154). However, the major part of the deposits are probably Quaternary although locally they may be as old as Cretaceous.

Alluvium

Stream deposits, which comprise only a very small portion of the area studied, are usually masked by colluvial deposits and thus, only alluvium related to the larger streams is mapped separately.

The alluvium associated with the southern branch of Conewago Creek near the eastern margin of the map is a restricted deposit comprised of gravels and sand derived from the rocks of the Catoctin Formation and the lower units of the Chilhowee group. The matrix is primarily clay.

Conococheague Creek has developed the largest alluvial deposit. It begins near Caledonia Park and extends westward, reaching maximum development near the western margin of the area where the creek flows on weaker rocks. Gravel, sand, and clay are the primary components of the alluvium. The coarser particles can generally be recognized as fragments of the Antietam and Montalto units, both of which are exposed in adjacent ridges.

METAMORPHISM

Metamorphism in the study area is regional (Turner and Verhoogen, 1951, p. 371) and unrelated to any known igneous activity. Conversion of the calcic plagioclase and pyroxene of the Catoctin basalts to albite-epidote-actinolite, and the quartz-microcline-albite-chlorite mineralogy of the upper Harpers, an original pelitic or silty graywacke, identify the greenschist facies of metamorphism as defined by Turner and Verhoogen (1951, p. 456-473). The association of muscovite-sericite with chlorite and quartz plus the absence of biotite in the quartz-bearing pelitic rocks suggest that the rocks represent the muscovite-chlorite subfacies of the greenschist facies (Turner and Verhoogen, 1951, p. 469-472).

Notable compositional change is restricted to the volcanic and more argillaceous rocks; however the abundance of sericite in the matrix of the sandstones is probably the result of recrystallization of an original argillaceous matrix. Farlekas (1961, p. 40) indicates that there has been no mineralogical change in the Tomstown Dolomite. Only recrystallization of calcite and dolomite is reported locally in the other Cambro-Ordovician carbonates by Gair (1950, p. 868).

The effect of metamorphism is most strongly imposed upon the megascopic and microscopic fabric of the rocks. Metamorphism was accompanied by the development of a prominent cleavage in all but the most quartzose rocks. Additionally, flattening, extension, and the down-dip alignment due to rotation or growth of ooids, minerals, rock fragments, and other bodies in the plane of the cleavage is also developed.

In thin-section, dimensional alignment of grains and matrix occurs in most rocks. Fellows (1943, p. 1400, 1412-1414) recognizes various quartz types as evidence of deformation. These quartz types are present in the rocks of the Chilhowee Group of the study area and include peripheral growth quartz, crushed and needle quartz, and recrystallized quartz. The principal mechanical effects induced by the low grade metamorphism are intense undulose extinction, deformation lamellae, and peripheral granulation of the quartz grains.

STRUCTURAL GEOLOGY

INTRODUCTION

Eleven structural elements, including bedding, three cleavages, and seven lineations are identified, in addition to joints. The identification and description of these elements in the field have been complemented by laboratory study and compilation. Many of the data are presented in fabric (contour) diagrams. Fabric diagrams represent projections onto the lower half of a Schmidt equal-area net.

Sander (1930, reported in Cloos, 1946, p. 4-5) has suggested a terminology for distinguishing planar elements. The term S-plane (S-surface) is used to identify sets of parallel to nearly parallel planes of mechanical inhomogeneity in deformed rocks, and is purely a descriptive, non-genetic term. Similarly lineations are designated by the letter L.

The orientation of both planar and linear structural elements can be described in terms of three mutually perpendicular coordinate axes as defined by Sander and described by Cloos (1946, p. 6). The coordinate system is such that the *b* direction coincides with the fold axis, the *a* direction is perpendicular to *b* and in the direction of tectonic transport, and the *c* direction is perpendicular to the *ab* plane.

To distinguish between several planar and linear structures, numerical subscripts are commonly employed. The numerical value assigned generally corresponds to the order of development of the respective planar or linear element. In this report S_0 is bedding, S_1 is flow cleavage, S_2' is slip cleavage, and S_2 is fracture cleavage. Lineations are described by letter and numeral subscripts. These relate the linear element to the coordinate axes or the planar elements whose intersection determines the lineation. Thus, L_a is the lineation parallel to the *a* coordinate direction, and L_{ox1} is the lineation formed by the intersection of bedding (S_0) and flow cleavage (S_1).

Figure 7 illustrates the coordinate axes and their relationship to the more common planar and linear elements developed in the rocks of the Caledonia Park area.

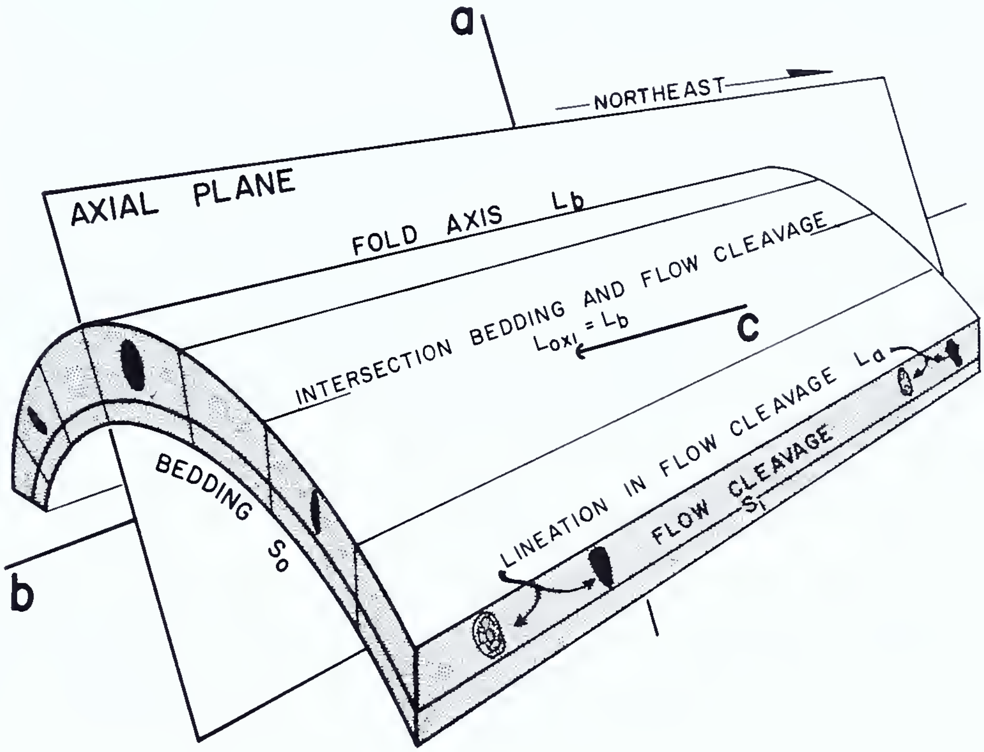


Figure 7. Schematic diagram showing the tectonic coordinates and major structural elements in an anticline.

STRUCTURAL ELEMENTS

Primary Planar Structures

Bedding, the dominant planar structure, is identifiable in nearly all of the exposed formations. It is typically outlined by changes in lithic character in the thin-bedded sediments, and as color bands, silty or hematitic laminae, pebble zones, or argillaceous or dolomitic laminations or bands in the more massive beds.

Flow banding and amygdule zones are the primary structures identifiable in the Catoctin volcanics. Flow lines, generally recognized by color change, are restricted to the rhyolite, where they occur locally. Vesicular zones occur in both of the major volcanic types as well as the epidosite. The size of the exposures and their scarcity limits the usefulness of these structures for determination of bedding or flow contacts.

Skolithos tubes which occur in the Montalto and Antietam frequently identify bedding because they are perpendicular to it. The tubes are particularly useful for determining bedding where other criteria fail, such as in the massive, well-jointed protoquartzites and quartzites.

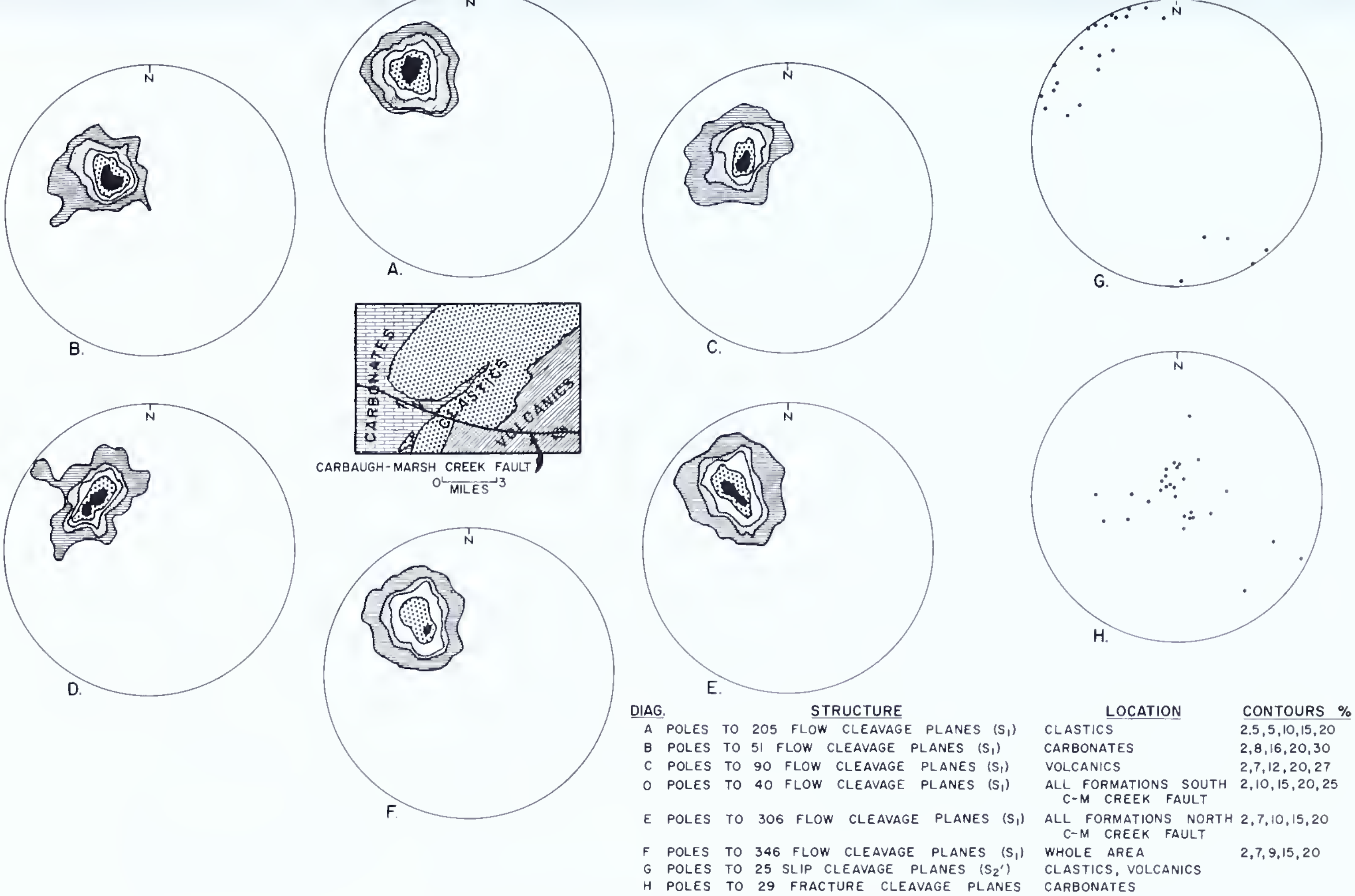
Close examination of all exposures usually reveals bedding even in rocks dominated by cleavage. However, locally, bedding in the Loudoun, Weverton, and Harpers Formations has been completely obliterated by cleavage development.

Many primary structures used in determining top and bottom are present in the sedimentary rocks of the region. Cross-bedding is a structure common to many beds in the Loudoun, Weverton, Harpers and Conococheague Formations. Graded bedding occurs locally within the Loudoun and Weverton Formations. It is usually best developed in the quartzose graywackes, and consists of pebbly or coarse grains which grade upward into fine to very-fine-grained sand over intervals of a few inches. Mudcracks, cut-and-fill structures, and stromatolites are confined to the carbonates and are best developed in the Conococheague Formation.

Secondary Planar Structures

Flow Cleavage (S_1)

The dominant secondary structure in the South Mountain area is a flow (regional axial plane) cleavage which is developed in nearly all formations. This cleavage (S_1) is the *ab* plane or plane of movement, and maintains a consistent northeast strike, moderate southeasterly dip, and position parallel to the axial planes of bedding folds (Plate 2, Figs. 7 and 8). The orientation of S_1 cleavage in the Caledonia Park area (Fig-



DIAG.	STRUCTURE	LOCATION	CONTOURS %
A	POLES TO 205 FLOW CLEAVAGE PLANES (S_1)	CLASTICS	2.5,5,10,15,20
B	POLES TO 51 FLOW CLEAVAGE PLANES (S_1)	CARBONATES	2,8,16,20,30
C	POLES TO 90 FLOW CLEAVAGE PLANES (S_1)	VOLCANICS	2,7,12,20,27
O	POLES TO 40 FLOW CLEAVAGE PLANES (S_1)	ALL FORMATIONS SOUTH C-M CREEK FAULT	2,10,15,20,25
E	POLES TO 306 FLOW CLEAVAGE PLANES (S_1)	ALL FORMATIONS NORTH C-M CREEK FAULT	2,7,10,15,20
F	POLES TO 346 FLOW CLEAVAGE PLANES (S_1)	WHOLE AREA	2,7,9,15,20
G	POLES TO 25 SLIP CLEAVAGE PLANES (S_2')	CLASTICS, VOLCANICS	
H	POLES TO 29 FRACTURE CLEAVAGE PLANES	CARBONATES	

Figure 8. Equal-area projections of cleavage.

ure 9) is in accord with similar data from Maryland and Virginia (Cloos, 1951, p. 144-146; Nickelsen, 1956, p. 258, Fig. 4a; Whitaker, 1955, p. 452, Figs. 4a, 4b).

Only minor differences in orientation of the flow cleavage occur in the major lithic domains (Figure 8, a-f). These differences are likely the effect of different structural position in the South Mountain fold as well as the physical properties of the rocks.

Locally this cleavage dips steeply northwest. This anomalous orientation may be the result of folding of the flow cleavage, diffierential slip of the cleaved bed between more competent beds involved in flexural slip folding, or local faulting.

The degree of development of the regional cleavage (S_1) varies with lithology. Although this structure is not confined to the Chilhowee Group,

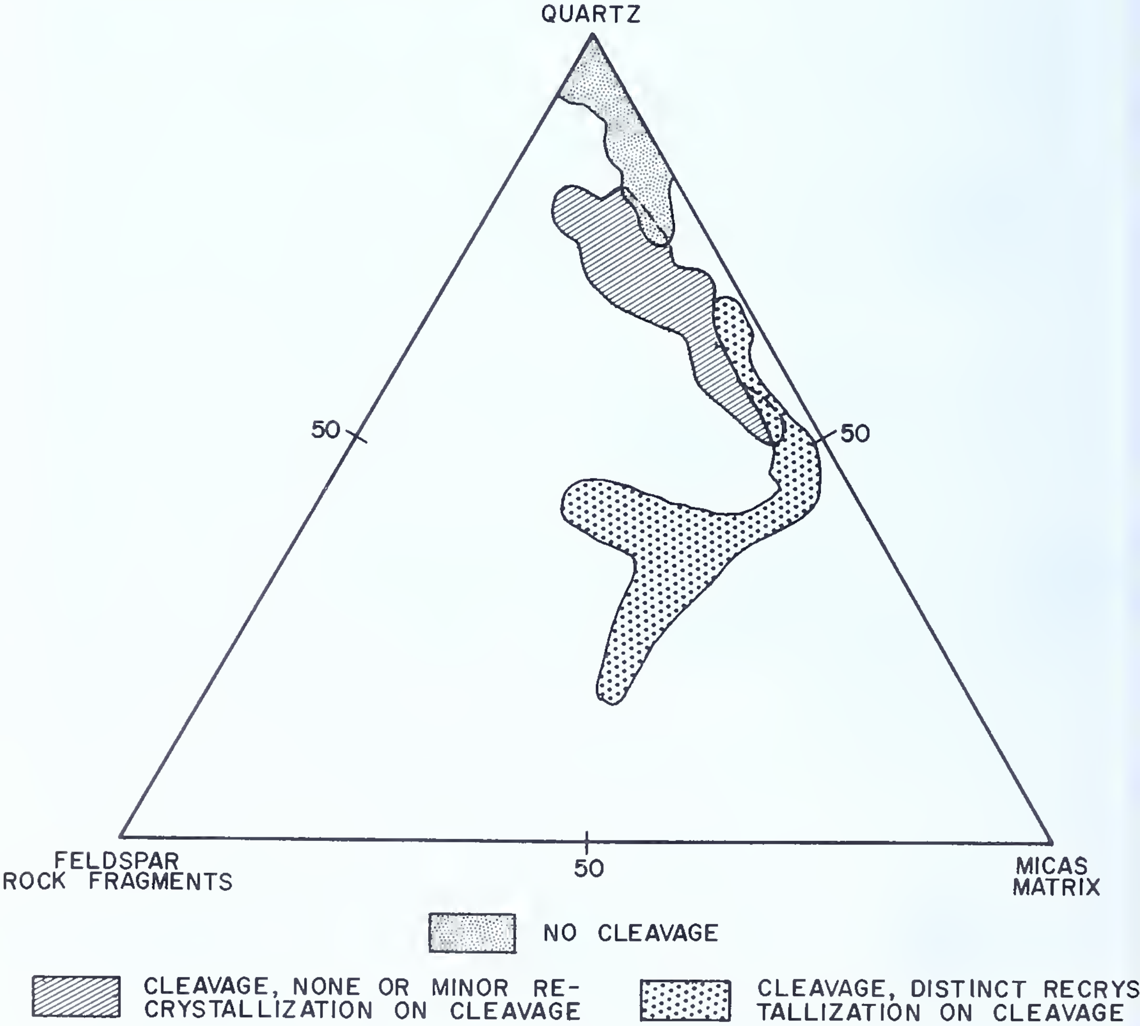


Figure 9. Relation between composition and flow cleavage development in the Chilhowee Group.

the relationship between the character of the cleavage and rock composition can be best examined in these rocks.

Cleavage planes are typically lustrous owing to recrystallization and orientation of micaceous minerals along these surfaces. Rocks with matrix in excess of about 30 percent (Figure 9) show megascopically visible micaceous flakes, some exhibiting slickensides down the dip of the cleavage. The more quartzose rocks show a cleavage consisting of non-lustrous, persistent partings which characteristically roughen the surfaces of exposures.

The S_1 cleavage not only reflects a parallel arrangement in inequant minerals, but also is the result of a shortening normal to the cleavage and a lengthening or movement parallel to this surface. Oolites in the carbonates and micropoikilitic patches in the rhyolites, both probably originally circular in outline, have been deformed into elliptical bodies lying in the plane of the cleavage (Plate 3, Figs 1-2). Similarly, clastic quartz grains have been altered to a lens-like form parallel to the cleavage trace through peripheral granulation and "tailing" (Plate 3, Figs. 3-4).

These features, and the slickensided micaceous material coating the cleavage surfaces, demonstrate that the development of S_1 cleavage involves more than mere relative movement of microlithons along fracture surfaces, but includes internal flow within the microlithons parallel to the fracture surfaces. Thus, it is properly called flow cleavage (Billings, 1954, p. 340).

Slip Cleavage (S_2')

Locally strata of the Weverton and Loudoun Formations, and the Catoctin phyllite unit are affected by a second cleavage which transects the regional flow cleavage (S_1). Characteristically this cleavage consists of a series of closely-spaced fractures or microfaults and is identified as slip cleavage (Billings, 1954, p. 339) or strain-slip cleavage (Turner and Weiss, 1963, p. 98).

The effect of the slip cleavage (S_2') is the disturbance of the earlier flow cleavage (S_1). This disruption is expressed as microfolding of the flow cleavage or a puckering caused by microfaulting in the axial region of the microfold. When deformation is not so intense as to create a microfault, the axial planes of the microfolds on the flow cleavage may be taken as the slip cleavage since these crinkles are the incipient stage of development of the slip cleavage (Nickelsen, 1956, p. 257).

The folds resulting from the disruption of the flow cleavage by slip cleavage are normally only an inch or two in amplitude. However, a few nearly a foot in height have been observed in float fragments (Plate 3, Figs. 5-6).

Microscopically, well-developed slip cleavage may show recrystallized sericite in the cleavage plane (S_2'). More often, a bending or shearing of the oriented micaceous minerals from the earlier flow cleavage has taken place.

Figure 8, g is a plot of poles to slip cleavage. Comparison of diagrams in Figure 8 shows the relationship between the normal southeasterly dipping flow cleavage (S_1) and slip cleavage (S_2'). The average angular difference in dip is 40 degrees; about 70 percent of the rocks exhibiting both cleavages show an angular discordance of 35 degrees. The data are in fair agreement with that of Nickelsen (1956, p. 257, Fig. 5, p. 260) which show an angular difference of 50 degrees in nearly 20 percent of the cases. The 15 degree average difference between the two localities is attributed to less steeply dipping flow cleavage in the Harpers Ferry area rather than any change in the slip cleavage attitude between that area and Caledonia Park.

The relationship between slip cleavage and bedding is not apparent owing to the scarcity of outcrops in which both these planar elements are identifiable. Nickelsen (1956, p. 257) reports no uniform relationship for the Harpers Ferry area between bedding and slip cleavage.

Fracture Cleavage (S_2)

A local cleavage is developed in some of the argillaceous and arenaceous limestones of the Elbrook and Conococheague Formations. It occurs as a series of sub-horizontal fractures spaced at intervals of a few millimeters or centimeters. As there is no evidence of mineral or particle alignment parallel to the cleavage surface and it is restricted to certain beds, the cleavage is interpreted as a fracture cleavage (Plate 3, Fig. 7).

In a few outcrops offset of clayey laminae is visible, indicating some movement along this planar surface. One exposure near Mountain Creek northeast of Scotland shows a small recumbent fold in association with this cleavage. The fold is a shear fold as interpreted from the thickening of bedding at the crests and thinning along the limbs. From this evidence, it seems that this fracture cleavage may grade to flow cleavage locally.

The attitude of the fracture cleavage (S_2) is clearly distinct from that of the regional flow cleavage (S_1) and slip cleavage (S_2') as shown in Figure 8, h.

Just south of the railroad along Conococheague Creek south of Woodstock, the sub-horizontal fracture cleavage (S_2) occurs in the Elbrook Formation. It is associated with locally developed concentric folds whose axial planes are nearly horizontal. Some of the adjacent unfolded beds exhibit only the regional flow cleavage (S_1). Although flow cleavage (S_1) and fracture cleavage (S_2) do not occur in the same rock, it seems quite certain that the fracture cleavage is a later cleavage developed in association with folds superimposed upon the regional fold style with its characteristic southeast-dipping axial plane and related flow cleavage.

The chronologic and genetic relations of slip cleavage (S_2') of the lower Chilhowee Group to the fracture cleavage (S_2) of the carbonates is unclear since they are restricted to nearly opposite positions in the stratigraphic sequence. That these two structural elements may represent the response of different rock types to the same stress condition is possible. However the writer interprets the fracture cleavage (S_2) as a response to localized gravity collapse affecting only the carbonates. This interpretation is consistent with the stress conditions required to form the recumbent bedding folds whose axial planes are paralleled by this cleavage. In contrast, the slip cleavage (S_2') is due to folding of the flow cleavage (S_1) and is a direct outgrowth of the stress field which produced the regional bedding folds and flow cleavage.

Joints

Joint surfaces representing joint sets not confined to a single bed were measured. The largest exposures furnished multiple groups of measurements.

Joint orientation is affected by the competency of the rock, and its stratigraphic and structural position. A single diagram of all joint readings usually does not yield a recognizable pattern, thus, diagrams were constructed for domains of similar lithology and structural character.

Various combinations of dip, strike, and oblique joints are recognized in the joint pole diagrams (Figure 10). Most of the diagrams fail to clearly define preferential orientations.

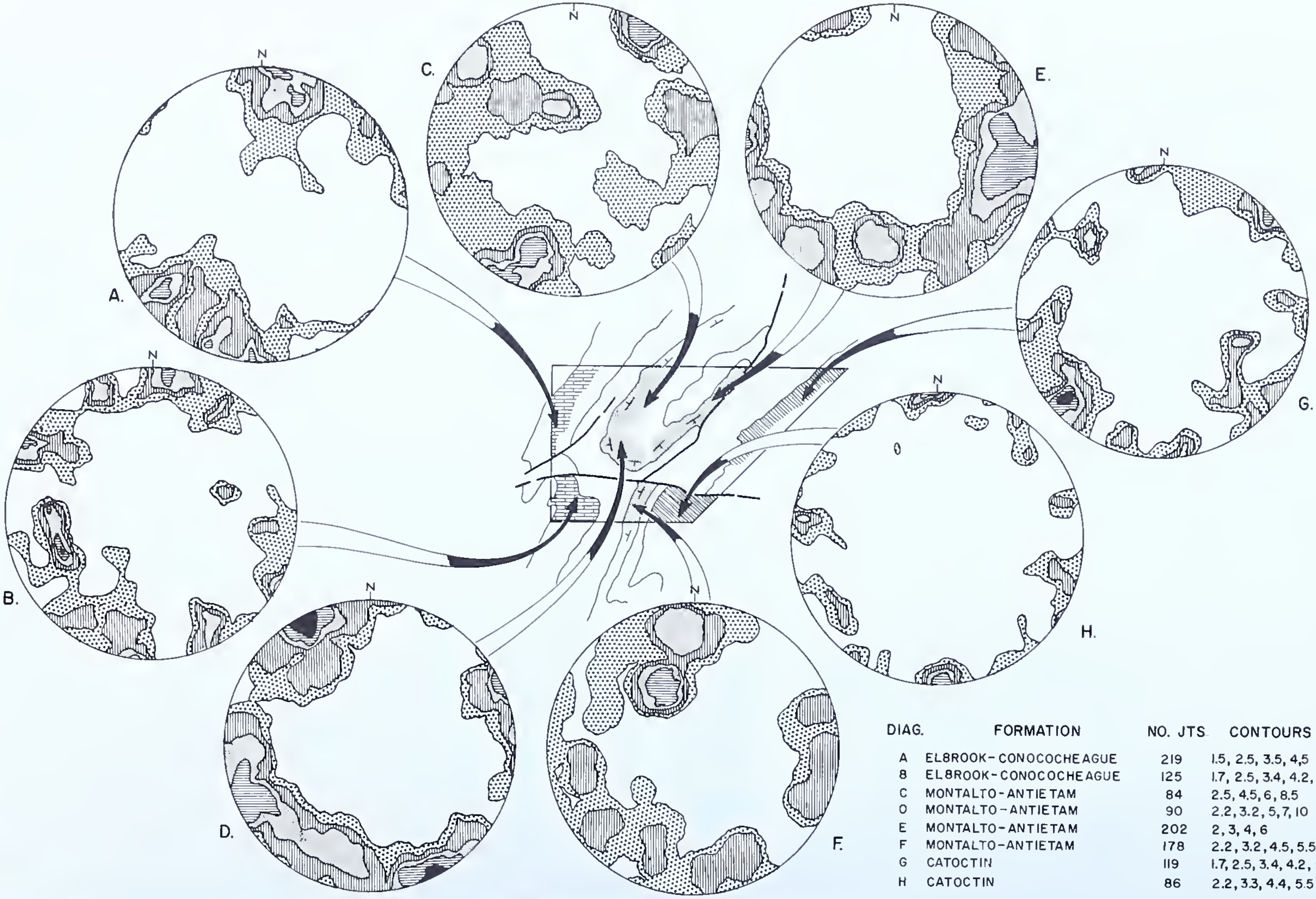
Joint pole patterns and maxima indicate the following:

(1) The majority of joints have a near-vertical attitude, reflected in the diagrams by an almost continuous girdle around the circumference. In general, the patterns are similar to that diagrammed by deSitter (1956, p. 132).

(2) There is a definite absence of joints with an attitude near that of bedding. This can be explained either by dissipation of the stress which produced the other joints through bedding plane slip, or by the development of joints sub-parallel to and indistinguishable from bedding.

(3) Joints oblique to the regional fold axes orientation commonly form pairs of maxima symmetrically oriented to the axes at angles of 25-35 degrees and 50-65 degrees. In each pair of maxima, one direction is more strongly developed than the other.

These maxima define conjugate joint orientations near those of theoretical surfaces of maximum potential shear, and thus represent shear joints. The position of these maxima is such that they are probably the result of a primary stress field (the stress field which produced the bedding



DIAG.	FORMATION	NO. JTS.	CONTOURS %
A	ELBROOK-CONOCOCHEAGUE	219	1.5, 2.5, 3.5, 4.5
B	ELBROOK-CONOCOCHEAGUE	125	1.7, 2.5, 3.4, 4.2, 5
C	MONTALTO-ANTIETAM	84	2.5, 4.5, 6, 8.5
D	MONTALTO-ANTIETAM	90	2.2, 3.2, 5, 7, 10
E	MONTALTO-ANTIETAM	202	2, 3, 4, 6
F	MONTALTO-ANTIETAM	178	2.2, 3.2, 4.5, 5.5
G	CATOCTIN	119	1.7, 2.5, 3.4, 4.2, 5
H	CATOCTIN	86	2.2, 3.3, 4.4, 5.5

Figure 10. Joint pole projections.

folds), and a secondary stress field. The secondary stress field is established as a result of the three-dimensional bending of the rocks during folding. Shear joints, such as those in Figure 10, d are symmetrical to, and interpreted as the result of, a local secondary stress system in which P_1 (major principal stress direction) was oriented northeast-southwest. This contrasts with the regional primary stress field with P_1 oriented northwest-southeast.

(4) Moderately eastward dipping joints occur in the steep western flanks of two major anticlinal structures (Figure 10 c and f). The position of their pole maxima closely coincides with the maximum for the regional flow cleavage (Figure 8). These joints represent one set of possible normal shear joints (deSitter, 1956, p. 132, Fig. 88), and may be the incipient stage of development of the regional cleavage (S_1) in the Antietam and Montalto sandstones and quartzites.

Linear Structures

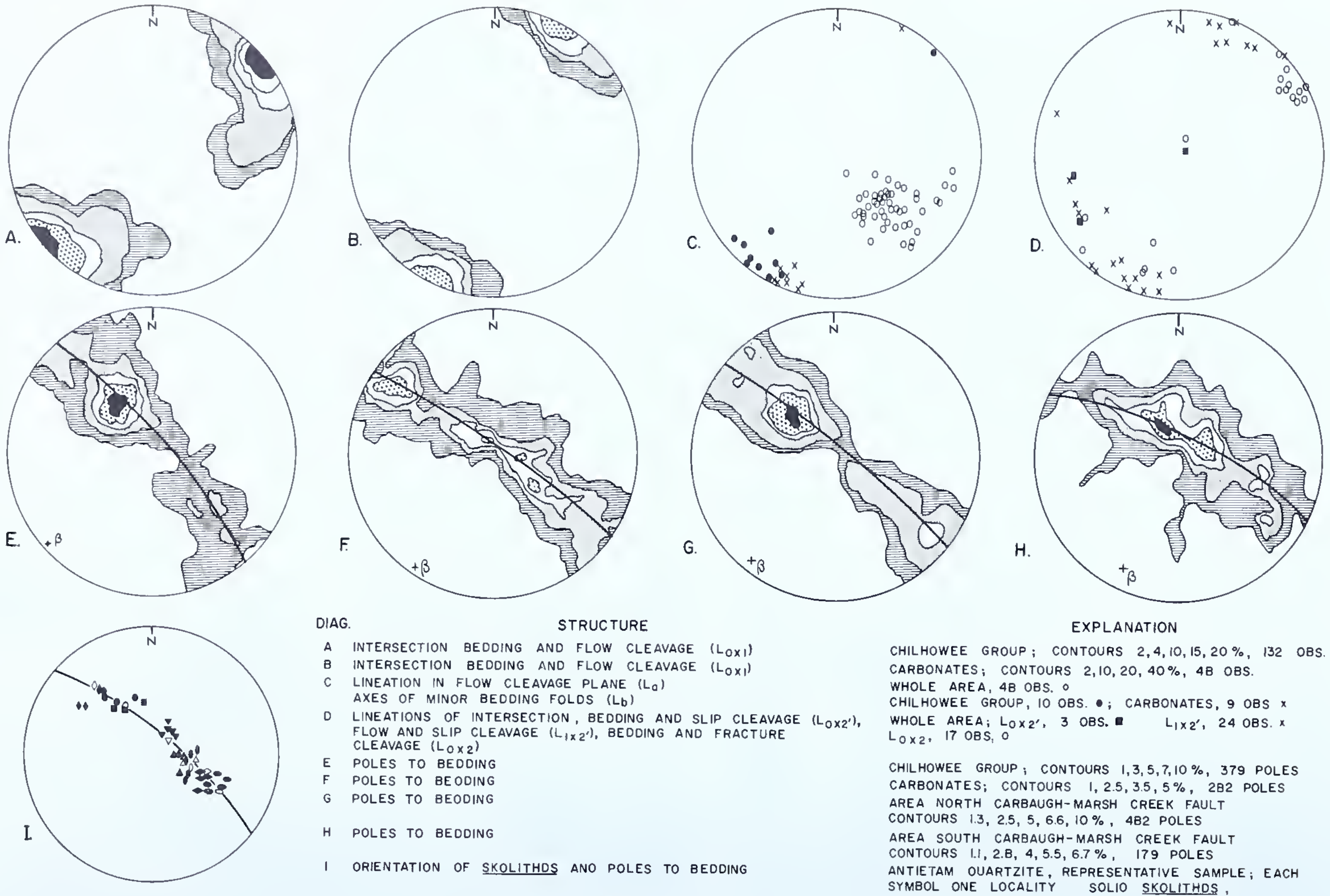
Lineations of Orientation and Elongation

Two classes of linear orientation are recognized. One is the orientation of *Skolithos* tubes, interpreted as worm tubes, and the second is the alignment of grains, rock fragments, and oolites. In undeformed rocks *Skolithos* tubes are cylindrical, parallel, and normal to the bedding. Measurements of the tubes, common in the sandstones of the Montalto and Antietam, show that they still maintain an attitude perpendicular to bedding throughout folded structures (Figure 11, Diagram i). The data support a flexural-slip mechanism for development of the folded *Skolithos*-bearing beds because pre-existing structures *(Skolithos* tubes) within a folded layer "maintain a constant angular relation to the attitude of the folded s-surface" in flexural-slip folding (Sander in Turner and Weiss, 1963, p. 473).

The characteristic lineation of the Caledonia area, and the South Mountain area in general (Cloos, 1957, p. 835), is the alignment of particles in the dip direction of the regional flow cleavage (S_1).

In the study area this lineation is visible in nearly all formations from the Catoctin volcanics through the Ordovician carbonates. Because of its consistent orientation in the cleavage plane (S_1) perpendicular to the fold axes, the lineation has been identified as an *a* lineation by Cloos (1946, p. 46, Plate 10; 1951, p. 146).

The *a* lineation is megascopically visible in the Catoctin volcanics as elongate chlorite blebs in the metabasalt and color streaks in the metarhyolites (Plate 3, Fig. 8). In the Loudoun phyllites, it is defined by paper-thin, highly elongate, green chlorite and ivory (sericitic?) streaks.



DIAG.	STRUCTURE	EXPLANATION
A	INTERSECTION BEDDING AND FLOW CLEAVAGE (L_{ox1})	CHILHOWEE GROUP; CONTOURS 2, 4, 10, 15, 20 %, 132 OBS.
B	INTERSECTION BEDDING AND FLOW CLEAVAGE (L_{ox1})	CARBONATES; CONTOURS 2, 10, 20, 40 %, 48 OBS.
C	LINEATION IN FLOW CLEAVAGE PLANE (L_a)	WHOLE AREA, 48 OBS. ○
	AXES OF MINOR BEDDING FOLDS (L_b)	CHILHOWEE GROUP, 10 OBS. ●; CARBONATES, 9 OBS x
D	LINEATIONS OF INTERSECTION, BEDDING AND SLIP CLEAVAGE ($L_{ox2'}$), FLOW AND SLIP CLEAVAGE ($L_{1x2'}$), BEDDING AND FRACTURE CLEAVAGE (L_{ox2})	WHOLE AREA; $L_{ox2'}$, 3 OBS. ■ $L_{1x2'}$, 24 OBS. x L_{ox2}, 17 OBS. ○
E	POLES TO BEDDING	CHILHOWEE GROUP; CONTOURS 1, 3, 5, 7, 10 %, 379 POLES
F	POLES TO BEDDING	CARBONATES; CONTOURS 1, 2.5, 3.5, 5 %, 282 POLES
G	POLES TO BEDDING	AREA NORTH CARBAUGH-MARSH CREEK FAULT CONTOURS 1.3, 2.5, 5, 6.6, 10 %, 482 POLES
H	POLES TO BEDDING	AREA SOUTH CARBAUGH-MARSH CREEK FAULT CONTOURS 1.1, 2.8, 4, 5.5, 6.7 %, 179 POLES
I	ORIENTATION OF SKOLITHOS AND POLES TO BEDDING	ANTIETAM QUARTZITE, REPRESENTATIVE SAMPLE; EACH SYMBOL ONE LOCALITY SOLID SKOLITHOS, OPEN BEDDING

Figure 11. Lineation diagrams.

Oriented quartz grains and quartzite, volcanic, and phyllitic pebbles occur in the sandstones and conglomerates of the Loudoun and Weverton. Additionally, cleavage surfaces coated with recrystallized micaceous material are marked by slickensides parallel to the *a* tectonic direction.

The *a* lineation is not commonly recognized in the quartzose rocks above the Weverton, although Cloos (1951, p. 146) has identified this structure in all the Lower Cambrian formations in Maryland. The *a* lineation reoccurs in the Cambrian and Ordovician carbonates as elongated oolites.

In thin section, this lineation is commonly expressed by elongation of micropoikilitic patches in the metarhyolites, of detrital rock and mineral particles in the lower Chilhowee strata, and of oolites in the carbonate rocks. Figure 11, c, developed from measurements made on oolites, color streaks and blebs, and rock and mineral fragments, shows the marked constancy in orientation of this structural element.

The oolites, originally spherical or near spherical bodies, have been distorted through deformation which affected the entire South Mountain region (Cloos, 1947, p. 897). Assuming no volume change during deformation, the percent of elongation in the tectonic directions can be easily determined. Two thin sections, cut to contain the three axes of the ellipsoidal oolite, permit the determination of the radius of the original sphere from which the ellipsoid has been derived. The volume of the sphere is equal to that of the ellipsoid; the relationship between the radius and semi-axes (half of the axes) of the deformed oolite is obtained from the equation:

$$4/3\pi r^3 = 4/3\pi a \cdot b \cdot c$$

$$r^3 = a \cdot b \cdot c$$

$$r = \sqrt[3]{a \cdot b \cdot c}$$

Percent elongation is determined by comparing the length of the semi-axes of the ellipsoid to the radius of the original sphere (Cloos, 1947, p. 862). Because the major axis of the ellipsoid lies down the dip of the flow cleavage and normal to the fold axes, and the intermediate axis is parallel to the fold axes, the axes of the ellipsoid parallel the tectonic coordinates.

Cloos (1947, p. 908-909, Plate 10) reports that elongation in the *a* tectonic direction (parallel to the major axis of the deformed oolite) reaches about 100 percent near the western edge of South Mountain and decreases to zero at a distance of 5-8 miles from the mountain front. The elongation in the *b* tectonic direction is normally less than 10 percent (Cloos, 1947, p. 888). Similar values were obtained by the writer for samples examined from the study area (Plate 3, Fig. 2).

Most sandstones and conglomerates in the Loudoun and Weverton Formations contain a variable number of elongated quartz grains. Peripheral granulation of the original grains resulting in the production of crushed quartz is the principle mechanism responsible for the elongate outline.

Granulation is concentrated in one portion or at opposite sides of the detrital grain. The crushed quartz formed is drawn out from the grain parallel both to *a* and the aligned micromicas of the matrix, resulting in a tapered "grain" which consists of a quartz grain and a streamer or "tail" of crushed quartz at one end or opposite ends of the core (Plate 3, Figs. 3,4). The crushed quartz "tail" is oriented parallel to *a* even where the longest diameter of the quartz core is transverse to the cleavage trace. Some quartz phenocrysts in the metarhyolites also show a similar structure.

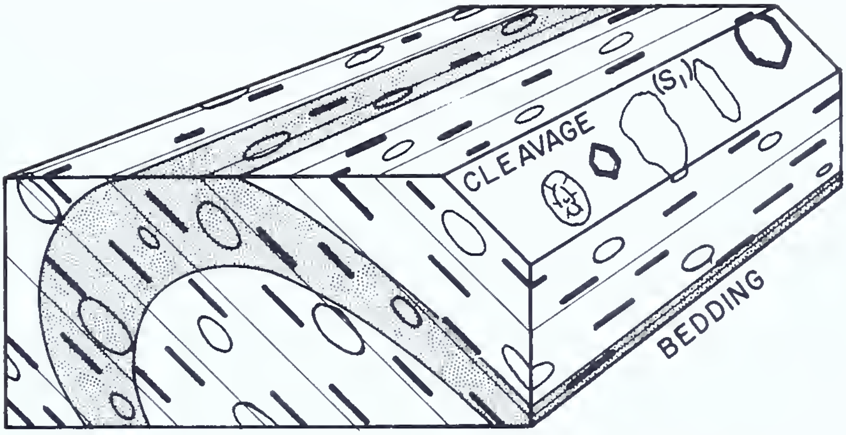
Normally estimated length-width ratios of the "tailed" quartz grains do not exceed about 2:1. Fellows (1943, p. 1411, Table 4; p. 1418, Table 5) reports ratios as high as 5:1 for some samples taken from the South Mountain area.

The groundmass texture of some metarhyolite is micropoikilitic. The micropoikilitic patches may have an equant to highly lenticular outline which is typically defined by a rim of very fine hematite or sericite, and optical homogeneity under crossed nicols (Plate 3, Fig. 1). The basic shape of these groundmass units remains nearly constant in a single sample. Examination of elongate micropoikilitic patches, whose length may exceed the width by three or four times, invariably shows that the major axis is subparallel to the dip of the cleavage in the hand specimen. Figure 8 h.

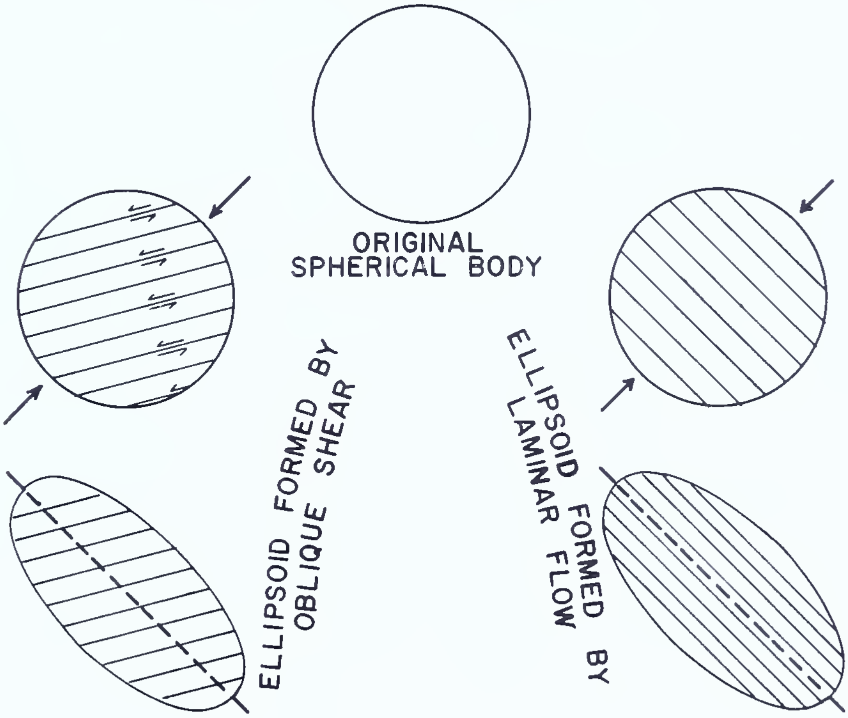
The significance of the *a* lineation is:

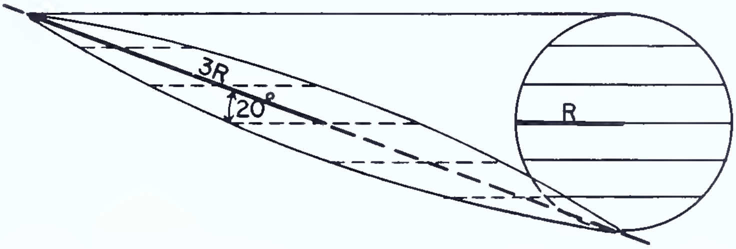
(1) Deformed oolites approximate a triaxial ellipsoid in which the major axis lies in the cleavage plane (S_1) parallel to *a*, and the minor axis is normal to the cleavage. This orientation between ellipsoid axes and cleavage is not compatible with the development of cleavage (S_1) by oblique shear.

If a sphere were deformed by shear along discrete, closely-spaced, subparallel fractures oriented at an acute angle to the compressive stress, the long axis of the ellipsoid formed would always be at an angle to the shear surfaces (Figure 12). This angle would measure about 20 degrees even if distortion reached 200 percent (Billings, 1950, p. 48). Since the measured elongation of the oolites is normally less than 200 percent (Cloos, 1947, p. 889, fig. 12), this angular deviation should be even greater. However, there is no measureable angle between the major axis of the ellipsoidal oolites and the dip direction of the cleavage (Cloos, 1947, p. 894-895). Thus, the regional flow cleavage is not caused by oblique shear.



LINEATION IN PLANE OF FLOW CLEAVAGE
MICAS (DASHES), OOLITES, AND ROCK FRAGMENTS (ELLIPSES) ORIENTED DOWN THE DIP OF CLEAVAGE PLANES


ORIGINAL SPHERICAL BODY
ELLIPSOID FORMED BY OBLIQUE SHEAR
ELLIPSOID FORMED BY LAMINAR FLOW



MAJOR AXIS OF ELLIPSE FORMED BY OBLIQUE SHEAR REMAINS INCLINED TO CLEAVAGE WITH DEFORMATION TO 300%

Figure 12. Development of flow cleavage (S_1).

(2) The relationship between the shape and orientation of the oolites and the flow cleavage (S_1) reflects a flattening process (deSitter, 1956, p. 214-215). Distortion is such that the axis of maximum elongation is normal to the axis of compression. The flow cleavage develops at right angles to the compression (Figure 12).

A similar mechanism is a valid explanation for the deformation of the micropoikilitic groundmass of some of the metarhyolites.

(3) The effects of flattening and elongation in mutually perpendicular directions may be visualized as unilateral slip along discrete surfaces spaced at submicroscopic intervals normal to the compression. This mechanism of rock deformation is essentially laminar flow. It is recognized and presented by Cloos as the principle process in the development of the regional cleavage (S_1) and the folding of the South Mountain region (Cloos, 1947, p. 908).

(4) Micro-structural features in rocks from the Chilhowee Group and Catoctin Formation suggest that stretching of the rocks took place parallel to the *a* tectonic direction.

Compression of a plastic body containing solid particles causes the dimensional alignment of the inequant solid grains. That non-aligned inequant grains may show crushed quartz "tails" invariably oriented in *a*, indicates relief from stress may also occur by granulation which may substitute for or precede dimensional alignment.

Some quartz grains are completely ruptured and separated. The intervening zone is partially filled by crushed quartz derived from the fragmented grain. Feldspar phenocrysts in some metarhyolites show a similar structure except that in this latter case, the intragranular space is filled by secondary quartz. The filled fractures are aligned nearly normal to *a*, and have the orientation of tension fractures.

The development of crushed quartz "tails" and the tension fractures in feldspar phenocrysts effectively cause the elongation of the particles parallel to *a*, and thus a stretching of the rock mass parallel to the flow cleavage plane.

The clastics of the Lower Chilhowee Group are primarily composed of two components, quartz grains and micaceous matrix (Figure 6). The components have unlike physical properties and thus respond differently to deformation. The more plastic matrix is likely to yield by laminar flow as the rock mass is compressed normal and extended parallel to the flow cleavage. The laminar flow of the matrix about the relatively brittle quartz grains could be the mechanism for attenuating the crushed quartz at the ends of the quartz grains parallel to the cleavage.

(5) Although the amount of distortion is variable, Cloos (1947, p. 856) indicates that elongation of ooids occurs in all samples taken from the general belt of oolite deformation. In contrast, aligned matrix, dimension-

ally oriented quartz, and crushed quartz "tails" are not invariably present in the rocks of the Loudoun and Weverton, and strained micropoikilitic patches do not characterize all of the metarhyolies with micropoikilitic groundmass.

For materials of different physical properties undergoing deformation, there is a distinct threshold value for each below which plastic or laminar flow does not develop. Thus, for a regional stress environment of given intensity, some rocks may deform by rock flowage while others do not.

If the carbonates were in a "soft" or semi-consolidated stage as Cloos (1947, p. 895-897) suggests, these rocks would be expected to show a lower threshold value and a greater tendency toward laminar flow than the older, and probably more highly consolidated sandstones and metarhyolites. This threshold requirement may explain the variation in intensity of the *a* lineation developed in the rocks of the South Mountain region.

Evidence of elongation in only some of the clastic and volcanic rocks may be the effect of variable local conditions. Variations in rock composition, pore fluids, structural position, and micro-stratigraphy are a few variables which may cause an intensification of stress conditions, or a decrease in the threshold limit locally. Either result could lead to a greater tendency toward rock flowage.

Fellows (1943, p. 1422) suggests that the clay content of a rock plays an important role in the way a rock responds to deformation. Cloos (1947, p. 880-881, figs. 9, 10) notes that oolite elongation varies with position on a fold.

It seems probable that a high matrix content could act effectively to absorb much of the stress or act as a buffer between grains. If so, the quartz grains in the rock should be less intensely strained. Less matrix would increase the amount of grain to grain contact and should favor adjustment through development of overgrowths by growth on unstressed and solution on stressed parts of quartz grains, intracrystalline slip, and crushing.

An examination of matrix content and quartz "tail" development fails to present any recognizable relationship (Figure 13, a). Similarly a plot of "tailed" quartz grains versus structural attitude yields no apparent correlation (Figure 13, b).

The lack of a correlation between quartz "tail" development and the two variables examined should not completely eliminate these factors from consideration in the analysis of the style of deformation. The results do imply that a more direct role may be played by such other factors as bed thickness, pore fluids, and local stratigraphic conditions.

(6) The *a* lineation defines the general direction of tectonic transport in the Caledonia Park area. This is in agreement with regional deformation plan of South Mountain (Cloos, 1957).

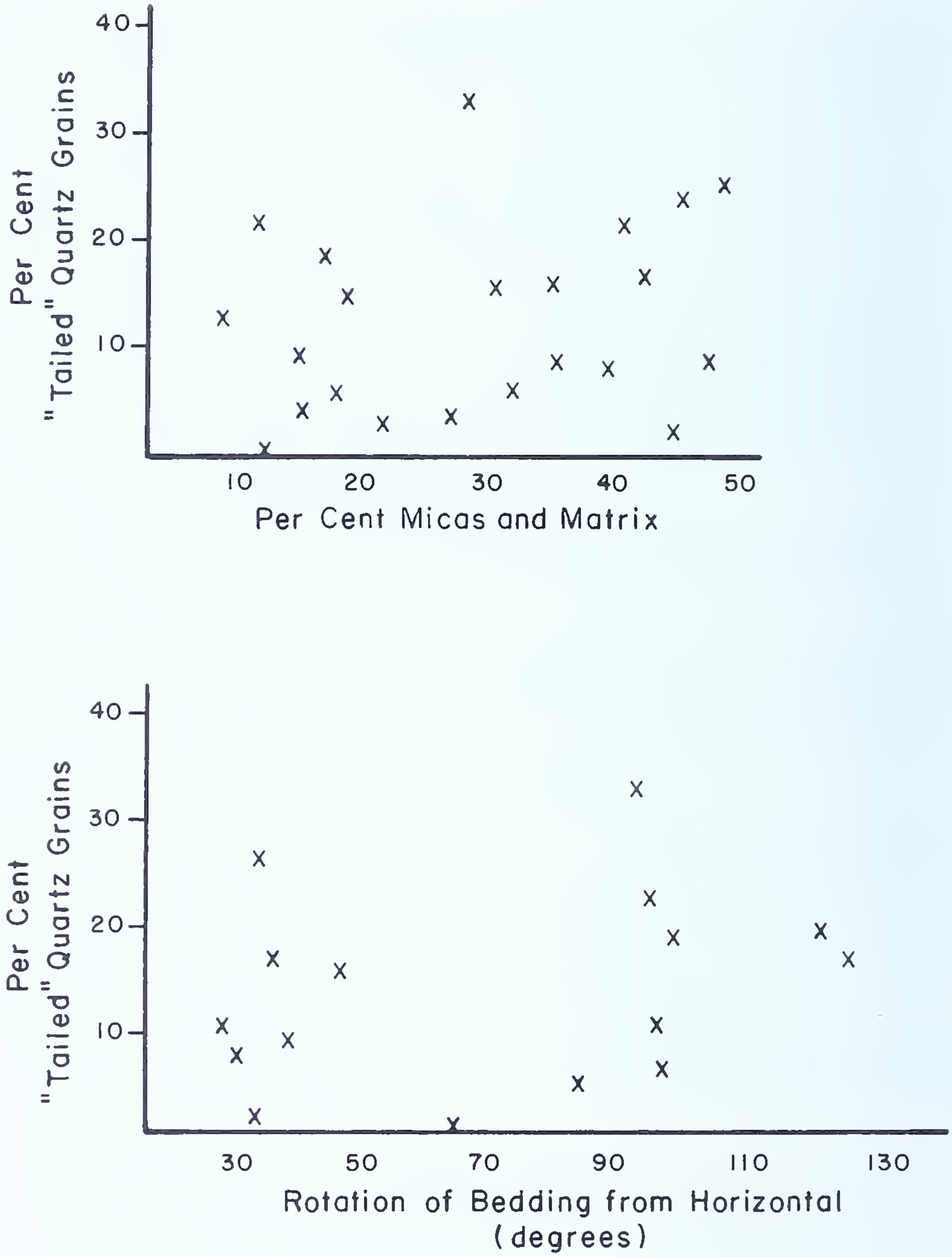


Figure 13. Relation of "tailed" quartz grains to rock composition and structural attitude.

Intersection of Planar Structures

Axes of small folds and lineations formed by the intersection of (1) bedding and flow cleavage (L_{ox1}), (2) bedding and slip cleavage

(L_{ox2}'), (3) bedding and fracture cleavage (L_{ox2}), and (4) flow and slip cleavage (L_{1x2}') comprise this group.

Although irregularities in the strike and plunge of bedding-flow cleavage intersections may result from the fanning of the cleavage (S_1), variation in rock competency, movement along the flow cleavage, and local structural dislocations, this element is the best developed and most consistently oriented lineation in the area. Figure 11, a, b shows that the lineation (L_{ox1}) very closely parallels the attitude of measured bedding fold axes (Figure 11, c). Consequently, this intersection is a *b* lineation and a valid estimator of the azimuth and plunge of bedding folds. This situation prevails in both the carbonate and clastic terranes.

The lineation defined by the intersection of flow cleavage (S_1) and slip cleavage (S_2') is limited to a few exposures of the phyllitic rocks of the Weverton, Loudoun and Catoctin Formations.

Slip cleavage (S_2') is typically associated with the crenulation of the flow cleavage (S_1) and results in the development of small chevron folds on the older flow cleavage (Plate 3, Fig. 5). The attitude of the axes of these cleavage folds can be measured directly, or, because the slip cleavage parallels the axial planes of the folds, can be determined graphically from the intersection of flow and slip cleavage.

Measurements of flow-slip cleavage intersections (L_{1x2}') and the axis of chevron folds reveal a low angle, northeast or southwest plunge for this lineation (Figure 12, d). The orientation is consistent with similar data from other parts of South Mountain (Nickelsen, 1956, p. 259; Whitaker, 1955, p. 433).

Unlike the bedding-flow cleavage intersection (L_{ox1}), this lineation (L_{1x2}') cannot be depended upon to outline the attitude of bedding fold axes even though it is not uncommonly subparallel to the trend of the regional folds.

The relationship of slip cleavage (S_2') to bedding and hence to *b* is not known since exposures are rare in which both elements can be recognized. In the three exposures where both are identifiable, the intersection (L_{ox2}') plunges to either the southwest or northeast, but the bearing and plunge exhibit marked variability (Figure 11, d).

The locally developed fracture cleavage (S_2) of the carbonates has an intersection with bedding (L_{ox2}) which coincides in orientation with the L_{ox1} lineation formed by bedding and the regional flow cleavage (Figure 11, Diagram c, d).

The relationship between the fracture cleavage (S_2) and flow cleavage (S_1) is unknown owing to the fact that both planar structures fail to occur in the same rock.

Folds

Two categories of folds are identified: folds on bedding and folds on flow cleavage.

The largest bedding fold is the Catoctin Mountain-Blue Ridge anticlinorium (Stose and Stose, 1946, p. 98) or the South Mountain fold or anticlinorium (Cloos, 1947, p. 853; 1951, p. 124). Bedding folds of various magnitude developed on the western limb of the anticlinorium are recognized in the study area (Figure 14).

Folds on flow cleavage are restricted to the phyllitic rocks in the Catoctin Formation and those which comprise the lower part of the Chilhowee Group. These structures are typically small chevron folds associated with the development of slip cleavage (S_2').

Folds on Bedding

Bedding folds are characterized by a moderately dipping normal limb on the east, and a steeply dipping or overturned limb on the west. Flow cleavage (S_1) approximately parallels the axial planes of the folds.

A plot of poles to a cylindrically folded S-surface defines a single great circle (Pi circle) on the stereonet. The normal to the Pi circle locates β, the long axis of the folded surface. Where the S-surface is bedding, β locates the fold axis (Turner and Weiss, 1963, p. 154-157).

Bedding measurements from the carbonate and clastic terranes are presented in separate, contoured Pi diagrams (Figure 11, e, f). The normal to the Pi circle of best fit for the data is an estimator of β, and hence the orientation of the axis of the bedding folds. The data, which are in accord with the measured axes of small folds and the intersection of bedding and flow cleavage (Figure 11, a, b, c), define a southwesterly plunge of 6 to 8 degrees.

Significant portions of bedding folds are exposed in only a few places. However, these outcrops provide some information on the character of the folded structure and the folding mechanism.

About 8 miles northeast of Caledonia Park, the axial region of an overturned syncline is exposed on the north side of Dead Woman Hollow between the elevations of 1580-1620 feet. The fold occurs in the upper part of the Weverton Formation which consists of thick-bedded, coarse-grained sandstones and thinner beds of phyllitic sandstone (Plate 4, Figs. 1 and 2).

Flow cleavage dips southeast at 50 degrees and is parallel to the axial plane of the fold. The plunge of the structure is about 5 degrees to the northeast.

The flow cleavage has a different orientation in beds of unlike competency in the overturned limb of the syncline. The cleavage which dips about 50 degrees in the thick-bedded sandstones, is inclined at 70-80 de-

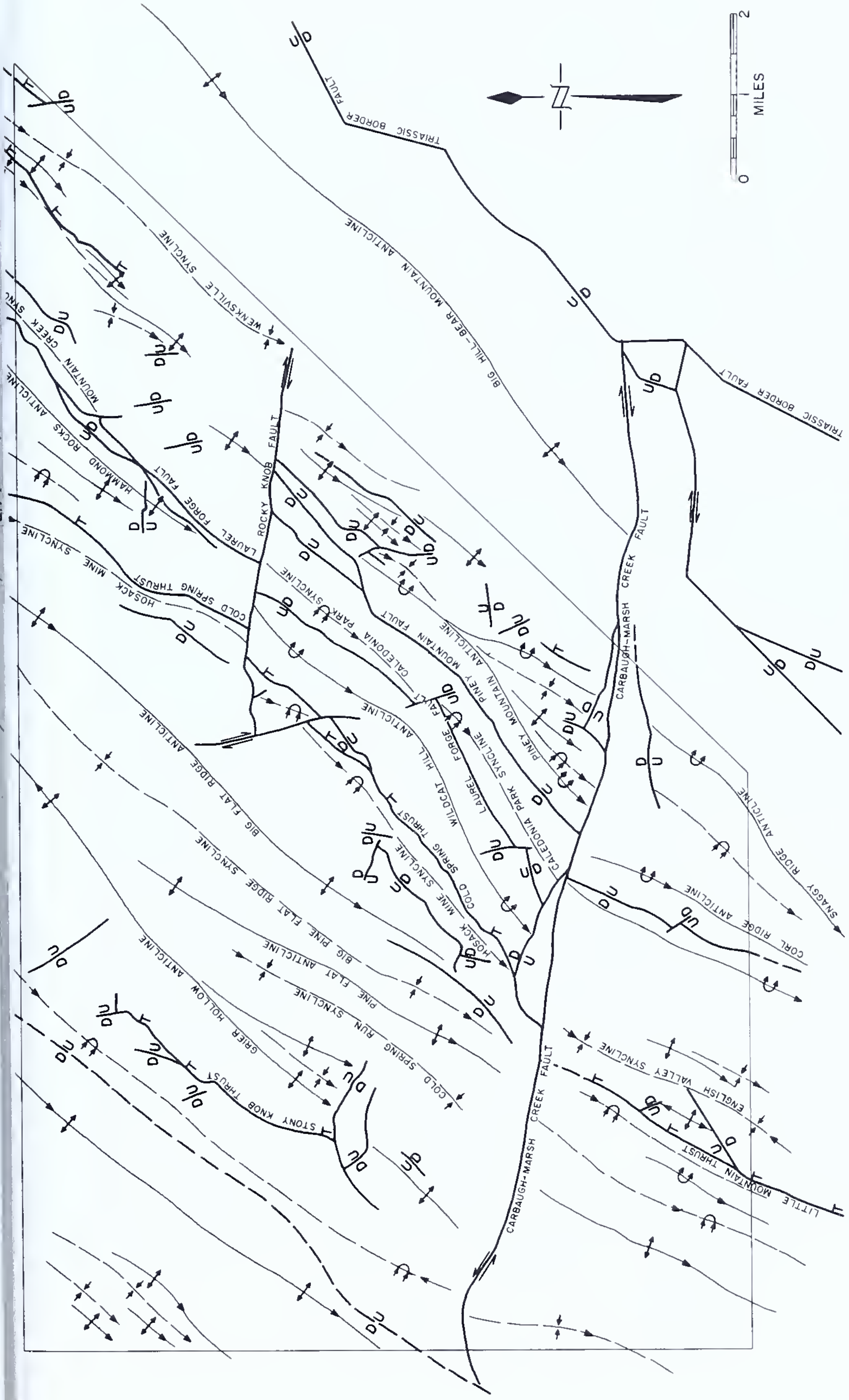


Figure 14. Fold axes and faults in the Caledonia Park area.

grees in the phyllitic interbeds. This marked change in attitude is usually considered to be the result of refraction across beds of different physical character. Although this change in orientation may be partially the result of refraction, an alternate mechanism which involves shear of the adjacent competent rocks across the incompetent bed is believed to play an important role. The sharp change in orientation of the cleavage, and drag at the contact which indicates the same movement sense that would be expected from differential movement of the more competent beds out of the synclinal trough, supports the second interpretation.

Relative thickening in the axial region of the fold, a prime characteristic of shear or similar folds, cannot be recognized because the limbs of the structure are not well exposed. However the lineated flow cleavage (S_1) and the microscopic character of the rock are evidence of important movement parallel to the cleavage plane. This suggests a shear mechanism for the development of the fold.

The structure in Dead Woman Hollow is typical of folds developed in rocks stratigraphically below the Montalto Member of the Harpers Formation. Shear folding is interpreted to be the dominant folding mechanism, but modification by flexural slip is recognized even in the later stages of folding. The degree of flexural-slip appears to be related to the relative competence of the rocks.

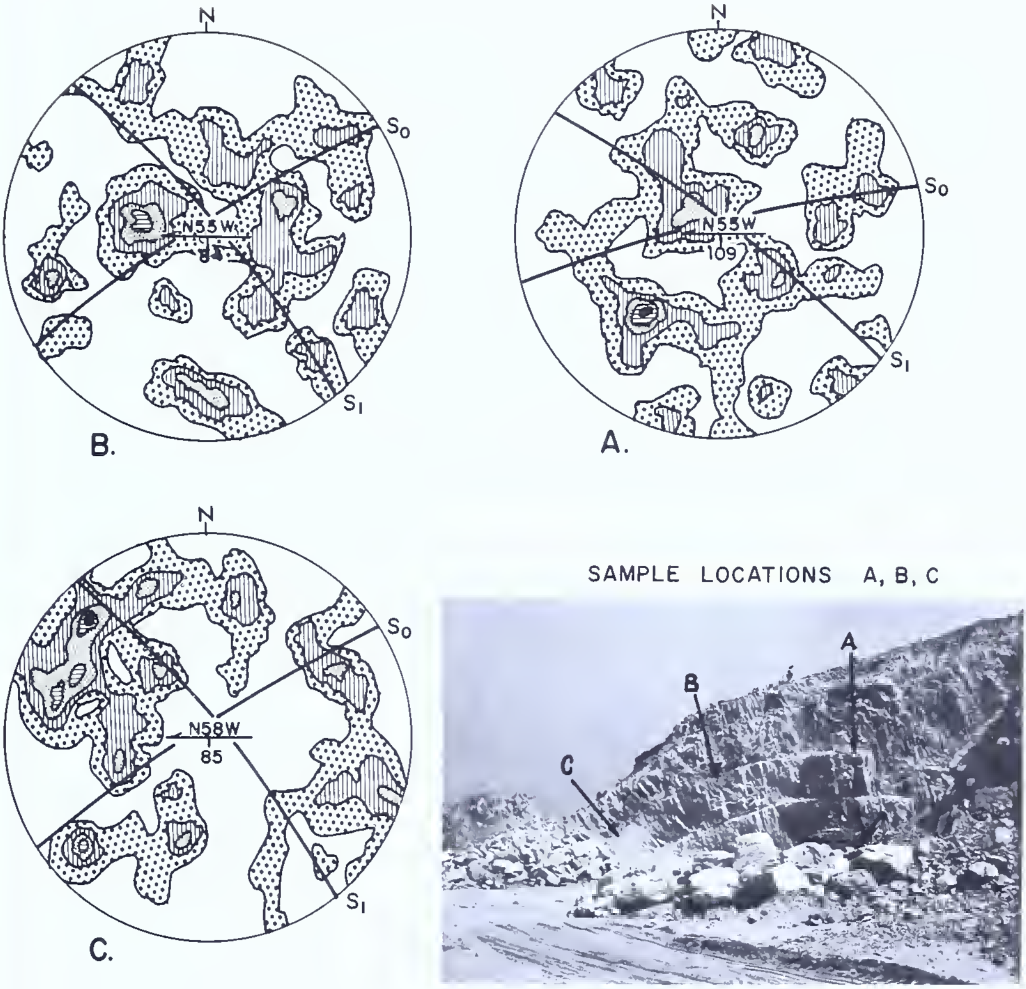
Segments of folds in the Antietam Formation are exposed in quarries located south of U. S. 30, two miles west southwest of Caledonia Park (Plate 4, Figs. 3, 4). The folds are asymmetrical and show a low southwesterly plunge.

These structures are developed in a stratigraphic section dominated by coarse-grained protoquartzites and quartzites. A few discontinuous beds of bluish clay, ½ - 3 inches thick, occur at irregular intervals.

Quartz grains in these rocks are equant to subsequant and exhibit intense undulose extinction, peripheral growth quartz, and deformation lamellae (Plate 4, Figs. 5-8). An extensive petrographic investigation of quartz-rich rocks of the Blue Ridge of Pennsylvania and Maryland led Fellows (1943, p. 1400) to conclude that "each (quartz) type indicates a particular degree of deformation, grading from original grains showing little or no deformation to completely reoriented and recrystallized SiO_2."

Deformation lamellae are the most prominent effect of intense deformation. Three samples taken from various positions on the western limb of an anticlinal structure show that up to 39 percent of the quartz grains exhibit these lamellae.

The orientation of the *c*-axes of quartz was also examined for these samples. The data are presented as fabric diagrams (Figure 15). The equal-area diagrams represent projections from the lower hemisphere.



DIAG.	BEDDING	NO. GRAINS	CONTOURS %
A	N55W 19W	200	1, 2, 3, 4, 5
B	N19E 35W	200	1, 2, 3, 4
C	N36E 52W	200	1, 2, 3, 4,

Figure 15. Quartz fabric diagrams.

Diagrams are oriented nearly perpendicular to the fold axis, thus the equatorial plane of the hemisphere is nearly vertical.

The attitude of the projection plane is given on each diagram. The dip angle is the angle defined by the viewer's line of sight and the plane of the projection; angles less than 90 degrees indicate planes dipping toward the observer, and angles greater than 90 degrees, planes dipping away from the observer. In all cases the viewer is facing northeast.

The *c*-axis orientation of 200 quartz grains selected at equally spaced intervals in each specimen is shown in Figure 15. Each diagram contains the trace of bedding (S_0) and, since flow cleavage is not developed in these rocks, a trace representing a possible cleavage orientation based on the measured plunge of the fold and the flow cleavage orientation in adjacent areas.

Two samples show maxima of 5 percent (Figure 15, a, c). In all cases concentrations of 4 and 5 percent are very limited and enclose less than 1 percent of the area of the net. The location of the maxima are unrelated to bedding or the regional foliation (S_1) except for Specimen A (Figure 15, a). In this sample point concentrations show one crystallographic *c*-axes orientation that is nearly parallel to the "potential" cleavage trace and another nearly normal to this plane.

Gair (1950, p. 870) notes the correlation of strong cleavage with distinct orientation patterns and the converse of this, and thus suggests that grains become oriented by movements along rock cleavage surfaces. If this relationship is assumed to be corrected, the pattern of Figure 15 a may be interpreted as tectonic.

Fellows (1953, p. 1425) accepts rocks with five per cent maxima as tectonites if advanced stages of recrystallization are present. However, it appears unlikely that the pattern of quartz axes obtained for rocks from this structure in the Caledonia Park area are of tectonic significance.

The folds exposed in the Antietam quarries are concentric folds produced by flexural-slip along bedding planes because:

(1) There is no systematic variation in thickness of the sandstone or clay beds across the structure.

(2) The regional cleavage (S_1) is not an identifiable structure in these rocks.

(3) *Skolithos* tubes maintain a consistent orientation normal to bedding regardless of the attitude of the strata.

Because the Antietam and Montalto are lithologically similar, the folds in these units are interpreted as concentric folds produced by flexural slip along bedding.

The major bedding folds in the carbonates are asymmetrical or overturned to the west. Flow cleavage approximates the axial plane of the folds and dips southeast. Deformed oolites define a prominent lineation which parallels the dip of the cleavage (S_1).

The bedding pole or Pi diagrams (Figure 11, e, f) and the intersection of cleavage (S_1) and bedding (S_0) (Figure 11, c) shows that the folds plunge at low angles to the northeast or southwest.

Cloos' analysis of the folds in the South Mountain Cambro-Ordovician carbonate sequence is based, in part, on exposures and data from the area of this study (1947, p. 901-911). An independent examination of many of the same outcrops by the writer corroborates Cloos' interpretation of the fold mechanism.

"The South Mountain fold is an excellent example of a large shear fold with laminar flow . . . .

The deformed ooids prove that the flow planes are cleavage planes and that considerable movement took place on cleavage. Laminar flow is the principal mechanism of folding and ooid deformation, for it is movement on subparallel planes and thus of equivalent effect as slip in many closely-spaced subparallel shear planes." (Cloos, 1947, p. 908).

An old limestone quarry and an exposure south of the railroad along Conococheague Creek, about five miles west northwest of Caledonia Park, exhibits folds of anomalous orientation with respect to the regional fold geometry.

The structures developed in some of the dense, fine-grained and shaly limestones of the Elbrook Formation are isoclinal, recumbent folds having an amplitude of 4-6 feet and wave length of less than 12 feet. Although the plunge of these folds is consistent with the regional fold axes, the axial plane is sub-horizontal. A locally developed cleavage (S_2) occurs parallel to the axial plane in the more shaly limestone units (Plate 5, Fig. 1).

The folds are not developed in all of the exposed strata. Some beds have a uniform sub-vertical attitude and exhibit only the regional flow cleavage (S_1).

Because the sub-horizontal cleavage (S_2) is restricted to the shaly limestones, has a variable attitude, and is only locally observed to offset bedding, it is interpreted as a fracture cleavage.

Minor folds developed in the cleaved argillaceous limestones at the Conococheague Creek locality, and also just east of Scotland, generally show thickened axial regions and a constant distance between bedding planes measured parallel to the cleavage (S_2). These structures exhibit the geometry of similar folds and are interpreted to be the result of shear folding (Plate 5, Fig. 2).

However, the massive, non-cleaved limestone at Conococheague Creek is characterized by recumbent concentric folds. These folds provide a basis for determining the amount of shortening represented by these folds. Field measurements show 15-20 percent shortening parallel to bedding. If the measured shortening affected all beds uniformly, the unfolded, sub-vertical beds probably have been thickened normal to bedding through recrystallization under the influence of a vertically oriented P_1.

Except for flexural-slip folding in the highly competent rocks of the Antietam and Montalto formations, the fold mechanism which characterizes the Caledonia Park area is shear folding. Folds may have started as flexure folds produced by bending and bedding-plane slip, but as folding continued slippage along cleavage planes was initiated and became progressively dominant. The lineated regional flow cleavage (S_1) was produced by plastic laminar flow, possibly aided by shear parallel to the cleavage. However in the later stages of folding, flexural-slip along bedding was still active in many rocks and probably produced the marked difference in orientation of the flow cleavage (S_1) in rocks of different competency.

Toward the end of the period of folding, small, recumbent flexure and shear folds developed locally in the carbonates. These folds appear to be "gravity" structures caused by the collapse of some beds located in the overturned or vertical limb of the older shear folds.

Folds on cleavage

Folded cleavage occurs mainly in the eastern portion of the area and only in the phyllitic beds of the Catoctin, Loudoun, and Weverton Formations.

The small, symmetrical chevron folds whose amplitude and wave length are measured in inches are developed on the regional flow cleavage (Plate 3, Figs. 5, 6). Similar folds are described by Nickelsen (1956, p. 261-63) and Whitaker (1955a, p. 454) from other parts of South Mountain.

In many cases the corrugations on the flow cleavage (S_1) are associated with slip cleavage (S_2'). This second cleavage (S_2') occurs at intervals of a few inches or less, and is oriented at a high angle to the flow cleavage (S_1) and parallel to the axial plane of the cleavage fold. Flow cleavage (S_1) is commonly offset by the other cleavage (S_2').

The lineation formed by the intersection of flow cleavage and slip cleavage (L_{1x2}'), and the axes of cleavage folds are sub-parallel to the regional bedding fold axes. This parallelism of L_{0x1} and L_{1x2}' is evidence that the bedding and cleavage folds resulted from a stress field of similar orientation.

Another lineation, in the form of striations, lies in the flow cleavage plane at a large angle to the axes of the chevron folds. This lineation maintains the same orientation on opposite sides of the chevron folds and is also developed on quartz layers that were injected as veins along the flow cleavage (S_1) subsequent to the development of these cleavage folds. The lineation resulted from the bending of the flow cleavage (S_1) and subsequent slip along these surfaces on the flanks of the cleavage plications.

The slip cleavage (S_2') is not developed in the quartz-rich layers or in the vein quartz. However in the micaceous zones, it locally disrupts the alignment of the micas parallel to the flow cleavage (S_1). Thin sections show the micaceous material "dragged" into the plane of the slip cleavage (Plate 5, Figs. 3, 4).

The development of the cleavage folds is interpreted as follows:

(1) Flow cleavage planes were folded by flexure with slip occurring between flow cleavage planes.

(2) Later slip cleavage (S_2') developed in the axial portion of the plications.

(3) Further development of the chevron folds continued by a combination of flexural slip along flow cleavage (S_1) in the more competent beds and by movement on the slip cleavage (S_2') and flow cleavage (S_1) in the less competent intervals.

Faults

The remapping of portions of South Mountain in Maryland and northern Virginia reveals few major faults, and folding is proposed as the dominant mechanism of deformation (Nickelsen, 1956; Cloos, 1951, 1957).

In contrast to the areas farther south, a recent study by the writer (Fauth, 1962) of the Pennsylvania segment of South Mountain shows a greater development of faulting than previously mapped (Stose, 1909, 1929, 1932).

Faults are not easily recognized in the Caledonia Park area because:

(1) The thickness of mappable units is large.

(2) Similar lithologies reoccur throughout much of the stratigraphic section.

(3) Stratigraphic control is limited; details of the lithic sequence and reliable measurements of thickness are generally unobtainable.

(4) Formational contacts are gradational.

(5) Bedrock exposure is limited.

In the present study, the use of aerial photographs and informal mapping units of distinct lithology and limited thickness form the basis for the recognition of faults. The location of the major faults is shown in Figure 14.

Two major faults, the Carbaugh-Marsh Creek and Rocky Knob faults, are transverse structures. The Carbaugh-Marsh Creek fault is located in the east-west valley occupied by Marsh, Carbaugh, and Conococheague Creeks. It is a distinguishable structure from the eastern border of the map area to the vicinity of Mt. Union, a distance of about four miles. West of Mt. Union the fault enters an area of thick, surficial gravel deposits and is not traceable.

A fault of similar orientation and movement to the Carbaugh-Marsh Creek structure is mapped just north of West Fayetteville. It is possible that these two faults are the eastern and western portions of the same structure. Such an interpretation is similar to that of Stose (1929, p. 13, fig. 11; 1932, p. 81, fig. 7).

The eastern portion of the Carbaugh-Marsh Creek fault juxtaposes Chilhowee Group rocks on Piney Mountain on the north and Catoctin volcanics on the south. The offset of formational contacts diminishes westward.

Important strike-slip movement is suspected for the Carbaugh-Marsh Creek fault because it:

(1) consistently offsets major structural axes uniformly westward on the south side

(2) separates two blocks of dissimilar fold development; folds are more numerous and more completely developed in the Chilhowee rocks to the north of the fault than to the south

(3) causes a distinct change in the topographic and structural trend of South Mountain (Plate 1; Figure 11, g, h)

(4) Maintains a straight trace across the topography, an indication of a steep-dipping fault plane

(5) abruptly terminates structural and topographic features.

If the interpretation of Carbaugh-Marsh Creek fault is correct, then it is a right lateral or dextral strike-slip fault. Horizontal displacement as determined from the offset of the Catoctin-Chilhowee contacts and fold axes is about 10,000 feet.

Important movement occurred along this fault during folding and after the development of the regional cleavage (S_1) as the flow cleavage (Figure 9d, c) and fold axes (Figure 12g, h) are rotated about 15-20 degrees counterclockwise south of the structure compared to the orientation in the northern block.

Several longitudinal faults north of the Carbaugh-Marsh Creek structure terminate against this fault, and their continuations have not been identified south of the fault. It is not imperative that a "post longitudinal fault" time be assigned as the age of the Carbaugh-Marsh Creek fault. The mapped relations are entirely consistent, and perhaps better explained, in terms of a pre- or early-folding time of formation of the Carbaugh-Marsh Creek fault and the subsequent independent deformation of the adjoining areas. Differences in the intensity of folding across this transverse fault give credance to the latter interpretation.

Another major transverse fault is located about five miles north of Caledonia Park. This structure begins between Rocky Knob and Sier Hill, passes eastward across East Big Flat Ridge and Piney Mountain, and enters the volcanic terrane about 1½ miles north of Strasbaugh School.

The Rocky Knob fault is truncated at its western extremity by a small fault. Together the two faults enclose an acute angle of nearly 60 degrees whose bisectrix is normal to the northeast-trending fold axes. The orientation of the faults approximates that of theoretical shear planes.

Major lateral movement along each fault is indicated by the eastward offset of both the Hosack Mine synclinal axis and beds of opposite dip. Horizontal offset is estimated at 2000-2500 feet for the Rocky Knob fault, and about 1000 feet for the fault to the south. These faults are apparently strike-slip and show right-lateral displacement. However, if the faults are a part of a conjugate shear set, both cannot show right-lateral displacement unless each developed and was active at different stages of deformation.

The Rocky Knob fault is probably a late-folding dislocation. This interpretation is supported by:

(1) Similarity in the character and degree of folding on opposite sides of the fault.

(2) Lateral movement along a plane whose orientation is a theoretical shear direction is a condition requiring a sub-horizontal position for the maximum (P_1) and minimum (P_3) stress axes. This necessary stress field orientation, with P_3 oriented northeast-southwest, can develop during the later stages of folding when stretching or extension parallel to the axes of folds, owing to the three-dimensional bending of the strata or the arcuate plan of South Mountain, is possible.

(3) The orientation of bedding-flow cleavage intersection (L_{ox1}) on opposite sides of the fault. This lineation trends S10°W on Rocky Knob in contrast to a trend of S40°W in adjacent outcrops. This shift in fold axis orientation, also reflected in the map pattern, necessitates a post-cleavage age movement on this structure.

Neither the Rocky Knob fault nor the fault immediately to the south along Knob Run was recognized by Stose (1929, 1932).

The remainder of the major faults parallel the structural grain and are thus longitudinal faults. These faults are identified mainly by the omission of strata. Supplemental evidence includes zones of brecciation and the termination or deflection of other structural elements.

Stose (1909, 1929, 1932) previously mapped many of the longitudinal faults, but failed to recognize the Piney Mountain, Rocky Mountain Creek, and Rocky Knob faults.

The attitudes of the fault planes are difficult to determine because they are not exposed. Straight fault traces may indicate a steeply inclined fault; however, low topographic relief along the strike of the structure rather than a steep fault plane may be responsible for the rectilinear fault traces. Topography does affect the trace of the Cold Spring thrust, and suggests a moderate southeasterly dip for this fault.

Minor faults are identifiable on the basis of mapped offset of stratigraphic contacts, or offset of topographic elements as observed on aerial photographs.

Minor oblique faults normally intersect fold axes at an acute angle; they parallel theoretically potential, primary shear planes. These faults have a straight trace and exhibit limited displacement and may be tear faults.

Brecciated sandstone is commonly found associated with minor longitudinal faults. The breccias are composed of poorly sorted, angular sandstone fragments set in a sandy clay matrix or cemented by hematite. The lack of stretching or flattening suggests that strong pressures normal to the fault plane were not present. Breccias of this type can develop under conditions of low confining pressure (shallow depth) or normal faulting.

Available evidence indicates that faulting accompanied the early and late stages of folding. The proximity of the area to the down-faulted Triassic basin and the parallelism of South Mountain and Triassic faults suggest that some of the normal faults may be Triassic in age. On similar evidence, Nickelsen (1956, p. 267) suggests this possibility for an east-dipping normal fault in South Mountain south of Short Hill near Hillsboro, Virginia.

In general, there is no direct evidence in the mapped area to date the faulting and folding other than that it was post-Cambro-Ordovician. The question to what extent the compressive structures reflect Taconic, Acadian, or Appalachian orogenic movements cannot be solved within the confines of this area.

REGIONAL GEOLOGY

The general stratigraphy of the Blue Ridge is a sequence of volcanics, clastics, and carbonates deposited in that order. Lithologic units occur in the Pennsylvania segment of South Mountain which are unlike those of comparable stratigraphic position in other areas as far south as Roanoke, Virginia.

The addition of rhyolitic rocks to the Catoctin Formation and the replacement of the graywacke siltstones of the Harpers Formation by protoquartzites and quartzites — the Montalto Quartzite Member — are the major modifications of the basic Blue Ridge stratigraphy in the Caledonia Park area.

Detailed correlations of Lower Cambrian lithologies and facies of south-central Pennsylvania and Maryland are impossible owing to limited bedrock exposure, the unfossiliferous nature of the rocks and the lack of detailed lithostratigraphic data. However certain gross regional relations are evident.

(1) The Loudoun Formation varies in thickness along the eastern margin of the South Mountain fold in Maryland. Near the southern end of Catoctin Mountain it pinches out and a banded quartzite of the Weverton overlies the Catoctin volcanics (Whitaker, 1955, p. 442-443).

In parts of York and Lancaster Counties, Pennsylvania, the Hellam Conglomerate Member of the Chickies Formation rests upon the Precambrian granitic surface. The Hellam is a quartz and jasper pebble graywacke conglomerate similar in character to the conglomerate unit of the Loudoun Formation and coarser beds in the Weverton Quartzite (Swartz, 1948, p. 1533). Based on reconnaissance studies east of South Mountain, the writer suggests that Loudoun and Weverton Formations thin markedly over a distance of 30 miles and are represented by the Hellam Conglomerate in western York County, a unit of similar lithology and stratigraphic position. This interpretation is similar to that of Swartz (1948, p. 1526, fig. 11).

(2) The lower three-fourths of the Harpers Formation, a graywacke siltstone, is gradually replaced northeastward in South Mountain by well-sorted protoquartzites and quartzites of the Montalto Member. On the basis of similarity in lithology and its stratigraphic position subjacent to the Harpers Formation, the Chickies Quartzite, a thinner quartzite body, is the correlative of the Montalto Quartzite east of South Mountain.

(3) The Antietam Quartzite is the oldest Cambrian formation containing useful fossils for age dating. It is a widespread, marine sandstone, which maintains a consistent lithologic character except for local areas of increased feldspar content in Virginia (Nickelsen, 1956, p. 251). This unit also thins eastward (Swartz, 1948, p. 1526, fig. 11).

(4) The Chilhowee Group changes in lithologic character from base to top. The general increase in grain size, quartz content, roundness, and sorting toward the top of the section in the Caledonia Park area is typical for much of the Blue Ridge belt north of Roanoke.

Graywackes and quartzose graywackes characterize the lower third of the Caledonia Park area sequence. These rocks grade upward through an increase in quartz content, and tectural and mineralogic maturity to protoquartzites and quartzites of the Montalto and Antietam. Generally feldspar content is low. A break in this pattern is represented by the upper Harpers which consists of less well-sorted graywackes siltstones than the adjacent strata, and is the only unit of the Chilhowee Group in Pennsylvania containing appreciable feldspar.

The Caledonia Park area is on the western limb of the South Mountain fold, a large structural unit with a uniform deformation plan. All structures are regionally oriented within the fold but vary locally with the competence and other physical properties of the rocks. All structures recognized in the area conform to this regional structural pattern and that described by other investigators in different sectors of South Mountain (Cloos, 1947, 1951; Nickelsen, 1956; Farlekas, 1961).

Bedding folds are asymmetrical and have steeply dipping or overturned western limbs. Flow cleavage (S_1) is parallel to the axial planes of folds and maintains a consistent southeasterly dip of about 45 degrees regardless of lithology.

Comparison of cleavage orientation northeastward from northern Virginia reveals a gradual steepening of the flow cleavage (Nickelsen, 1956, p. 258-259; Farlekas, 1961, p. 43-45) indicating that folds become more upright to the north.

Slip cleavage (S_2') is locally developed in the Caledonia Park area. The orientation of this planar element and its association with the folding of flow cleavage is a characteristic of the South Mountain anticlinorium. However, the stratigraphic range of this cleavage is more limited in the study area than in areas to the south, a reflection of the increasing competency of the stratigraphic sequence northward in the anticlinorium.

A locally developed cleavage in the Elbrook Formation appears to have no counterpart elsewhere in South Mountain. This cleavage (S_2) is an axial plane cleavage in local isoclinal, recumbent folds. It is interpreted as a structure formed penecontemporaneously with the folds; both the recumbent folds and associated cleavage (S_2) are the result of the gravitational collapse of certain beds in the Elbrook Formation on the steep western limb of the South Mountain fold.

An *a* lineation is recognized. This lineation is the most characteristic structural element present in the South Mountain fold as far south as the James River in Virginia. The *a* lineation retains its typical southeast dip in the plane of the flow cleavage (S_1) regardless of stratigraphic or structural position.

Pi diagrams of bedding in the Chilhowee Group and the Cambro-Ordovician carbonates reveal a southeasterly regional plunge of bedding folds. This is in accord with data compiled by Farlekas (1961, p. 50) for the Pennsylvania-Maryland border area. This plunge is contrary to that determined for northern Virginia by Nickelsen (1956, p. 260) and is opposite to the plunge near Mount Holly Springs (Stose, 1953). Thus, it is concluded that the axis of the South Mountain anticlinorium has a depression in Maryland, and a culmination north of the Caledonia Park area.

In general, mapping has shown that folding is not as intense as in areas to the south, and that faulting plays an important role in the deformation plan of this area.

The Carbaugh-Marsh Creek fault, a dextral strike-slip fault, causes a deflection in the orientation of structures near Caledonia Park. That it also separates blocks of dissimilar fold and fault development is particularly significant, as these structural relations appear to indicate that this fault developed early in the deformational history of the area.

GEOLOGIC HISTORY

In the absence of exposures of the Precambrian basement in south-central Pennsylvania, nothing is known concerning the sequence of events before the outflow of the Catoctin volcanics.

Prior to deposition of the volcanic and sedimentary rocks in the Caledonia Park area, a complex of granodiorite and biotite granite was formed. This terrane, now exposed in the core of South Mountain in Maryland and to the south, was deformed prior to sedimentation of the Swift Run Formation and outpourings of the Catoctin volcanics as shown by the divergence between the gneissic banding of the basement and the later regional flow cleavage of the South Mountain area. Subsequent to the deformation and erosion of the Precambrian gneisses, the volcanic and clastic sediments were deposited.

In southern Maryland, Virginia, and West Virginia the Catoctin basalts were extruded upon the eroded surface of the basement in the core of the South Mountain anticlinorium (Figure 16, A). These extrusives mostly covered areas of clastic sediments, mainly arkoses and graywackes containing fragments of granite and gneiss, which were deposited in depressions on the eroded surface (Reed, 1955, p. 877-878, 880-881). In other areas tholeitic basalts flowed out onto the eroded surface from feeders now represented by metadiabase dikes which cut the gneissic basement where exposed in Maryland (Reed, 1955, p. 890-891; Nickelsen, 1956, p. 245).

These Catoctin flows were subaerial (Reed, 1955, p. 894) as shown by the absence of pillow structures and the presence of columnar jointing. The rhyolite flows in the Caledonia Park region which are intimately associated with the basaltic rocks can only have formed in the same environment.

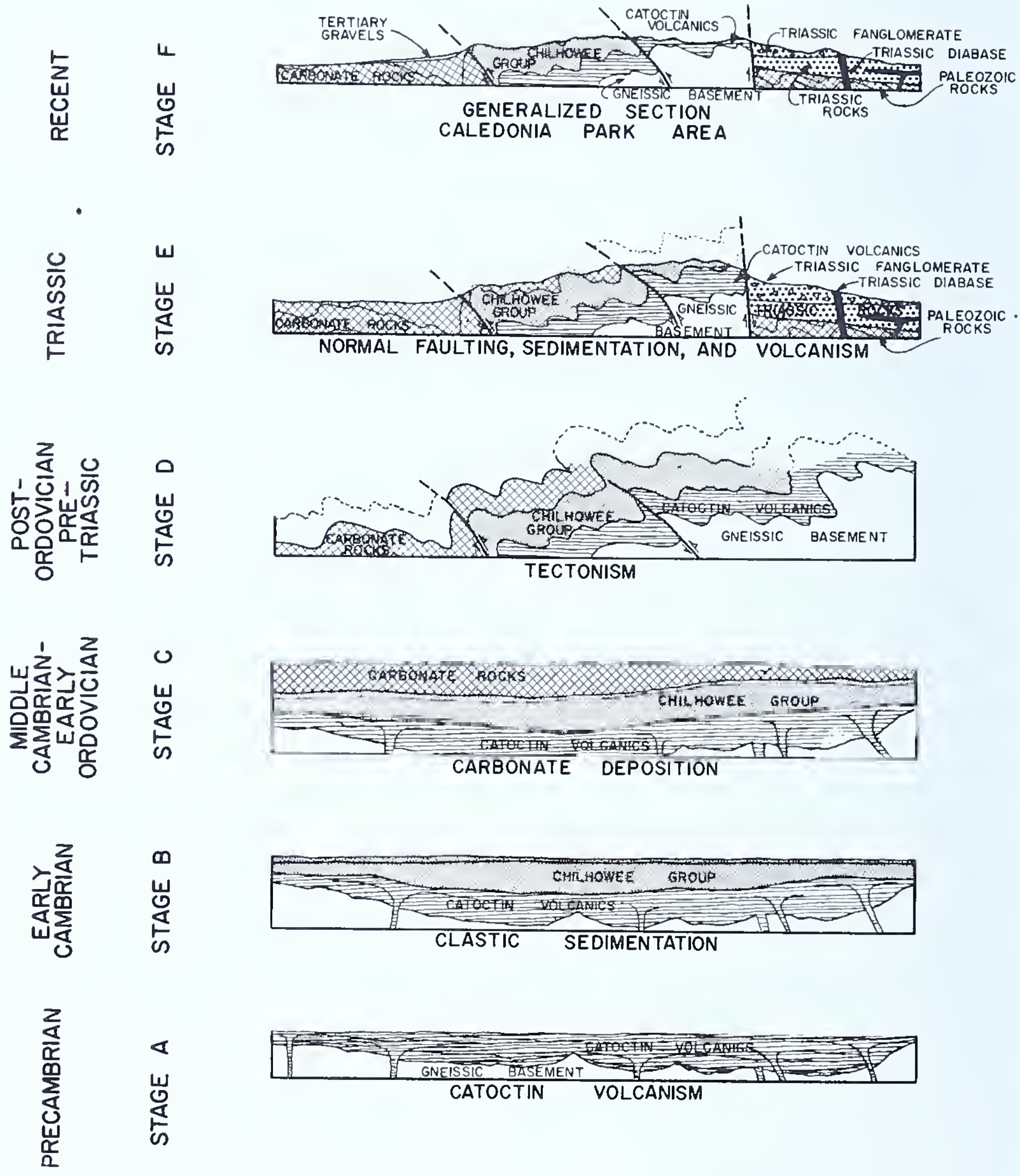


Figure 16. Geologic history of the Caledonia Park area.

The Early Cambrian sediments of the Chilhowee Group were deposited upon the Catoctin volcanics, or in some areas the gneissic basement (Reed, 1955, p. 878-879) during the waning stages of widespread volcanism (Figure 16, B). The rocks of the Loudoun Formation, the basal member of the Chilhowee Group, were formed from tuffaceous sediments (Nickelsen, 1956, p. 248) or saprolite (Reed, 1955, p. 891-893) and gravels derived from erosion of the diverse rock types already existing in the area. These rocks mark the transition from a dominantly terrestial volcanic environment to subaqueous clastic sedimentation, and represent the initial phase of a dominantly marine, Early Cambrian onlap under unstable shelf or marginal miogeosynclinal conditions.

Graywackes and quartzose graywackes, deposited conformably on the Loudoun, were gradually replaced by cleaner sandstones and quartzites. Increasing textural and mineralogic maturity in the Cambrian rocks reached its culmination with the deposition of the Antietam Quartzite, a widespread fossiliferous, marine unit of Lower Cambrian age.

The Loudoun and Weverton contain rhyolite fragments and plutonic quartz derived from a dominantly volcanic-granitic terrane affected by limited, low amplitude tectonic activity. The decrease in the thickness of these formations to the east has been interpreted as evidence for a southeasterly source for the detritus (Swartz, 1948, p. 1534).

However a source area to the north is likely for the Montalto because of the thickening of the sandstone facies of the Harpers Formation in that direction.

Decreased relief in the source area and increased tectonic stability in the area of sedimentation which caused a progressive shift from unstable to relatively stable marginal miogeosynclinal conditions, led to the replacement of low rank graywacke in the lower portion of the Chilhowee Group by protoquartzites and quartzites at higher stratigraphie positions during the Early Cambrian. The abundance and roundness of the stable minerals tourmaline and zircon, the well-sorted character of most of the sedimentary rocks, and fossil evidence (trilobite fragments and possible worm tubes) attest to a shallow water depositional environment within the miogeosyncline.

Tectonic or volcanic activity affected a limited area and caused a temporary interruption in these conditions. This resulted in abnormally high feldspar content and poorer sorting in the upper Harpers. The occurrence of a basic lava flow in the Harpers of the Sugarloaf area, Maryland (Scotford, 1951, p. 50), and tentatively identified rhyolite tuffs in the Catoctin Mountain area, Maryland (Whitaker, 1955, p. 446) support this interpretation.

Toward the end of the Early Cambrian the influx of clastic material decreased and the limestones and dolomites of the Tomstown Formation accumulated. The final event of the Early Cambrian was minor uplift in the source or a slow-down in basin subsidence which resulted in sedimentary up-building and the spread of clastics into the South Mountain area. This detritus is reflected in the sandy and argillaceous beds of the Waynesboro Formation.

Ripple marks, mud cracks, and other current features associated with the Waynesboro sediments resulted from shallow-water sedimentation. The location of the source lands for the Waynesboro detritus has not been established, but Swartz (1948, p. 1545-1549) suggests an area to the southeast on the basis of the absence of this formation in that direction.

Diminished relief in the source area during the remainder of Middle Cambrian lessened the influx of clastics. Under this setting extensive thin-bedded, argillaceous limestones and calcareous shales of the Elbrook Formation were deposited. In general, depositional areas in Pennsylvania during Middle Cambrian time were less extensive than in Early Cambrian and Late Cambrian-Early Ordovician time (Swartz, 1948, p. 1546). This condition probably resulted from a rate of sedimentation slightly in excess of basin subsidence which caused sedimentary up-building and the initiation of a minor regression of the shallow sea over part of the area.

The Late Cambrian Conococheague limestones were deposited in a transgressive shallow marine sea. The arenaceous beds at the base of the formation have been correlated with ". . . certain sand bodies of the dolomite-orthoquartzite facies . . ." (Gatesburg Formation) exposed in central Pennsylvania (Wilson, 1952, p. 317-318). The quartz content and grain size of these units increase westward toward a probable source in that direction (Swartz, 1946, p. 1050).

Cyclic carbonate sedimentation followed deposition of the basal Conococheague, and extended over a major portion of south-central Pennsylvania (Root, 1964, p. 157-160). Mechanical limestone in the lower part of the cycles formed at times when the depositional surface was above wave-base, whereas deepened waters of increased salinity favored precipitation of intercalated limestones and dolomites. The cycles are indicative of an offlap-onlap relation between environments of high and low mechanical energy (Root, 1964, p. 160). The large areal extent of the cyclic deposits reflect the shallowness of the Upper Cambrian sea, for under such conditions low amplitude oscillatory movements resulted in a major lateral shift in depositional environments.

Lithologically similar beds in the Conococheague and the basal portion of the Beekmantown Group, mainly represented by the Stonehenge Formation, resulted from continuance of Late Cambrian geographic and environmental conditions into the Early Ordovician.

Thus during much of Cambro-Ordovician time, the South Mountain region, the oscillatory margin of the miogeosyncline inundated by a shallow marine sea, was the site of carbonate sedimentation (Figure 16, C). Mild discontinuous or oscillatory movements in a source area of low relief periodically provided an influx of clay, silt, and sand. To maintain a shallow water character over a period of time in which thousands of feet of sediment accumulated,. slow downwarping of the basin was nearly in balance with sedimentary up-building.

Deformation of the rocks of the Caledonia Park area is related to a single period of tectonism which is divisible into several phases or pulses (Figure 16, D). A uniform tectonic plan, dominated by folding and the development of an associated regional flow cleavage and *a* lineation, was impressed upon all formations. All phases of the deformation were accompanied by faulting.

The age of the tectonism throughout the South Mountain region is interpreted as "post-Ordovician, at least Beekmantown" (Cloos, 1958, p. 20) on the basis that folding must be older than Triassic, which has no regional cleavage, but younger than those rocks affected by the South Mountain deformation plan. Thus deformation may be Taconic or Ordovician in age.

Radiometric age dates determined for regionally metamorphosed rocks from the Piedmont Province of southeastern Pennsylvania have been related to structural elements present in exposures along the Susquehanna River area near Lancaster (Lapham and Bassett, 1964). Although the rocks of the South Mountain region are not physically in contact with the dated rocks in the Piedmont, structures associated with the dated rocks are believed to correlate with the S_1 and S_2' structures in the study area. This correlation is based on the similarity in orientation, textural and structural characteristics, and the mutual relationship of the South Mountain planar elements to that of structures described by Freedman and others (1964) from rocks of the Lancaster area. If this correlation is valid, the suggested time of deformation in South Mountain is placed at about 330 million years (Lapham and Bassett, 1964, p. 665-666).

Dating of a mylonite zone associated with a bedding plane fault at the base of the Reedsville Shale (Upper Trenton age), folded in typical Valley and Ridge style, yields a value of 350 million years (Pierce, 1964). This structure is located about 20 miles west-southwest of Caledonia Park. The similarity in this age date with that obtained in the Piedmont strongly suggests that a Late Devonian time of deformation affected the South Mountain area — an event probably related to Acadian orogenic movements.

Deformation or at least the last phases of deformation may be the result of tectonism. Folded Pennsylvania strata in the Broad Top Basin and Southern Anthracite Field (40 miles west-northwest and 25 miles north-northeast of South Mountain respectively) which appear to show structural concordance with the dated rocks and their related structures (Freedman and others, 1964, p. 636) suggests this interpretation.

In Triassic time, South Mountain formed the western boundary of an elongate trough which was the locus of Triassic sedimentation (Figure 16, E). At first the sediments deposited in this trough, characteristically arkosic, were mainly derived from an uplifted granitic terrane located to the southeast. Later the axis of the basin shifted northwestward as that margin dropped along east-dipping normal faults. Relief on the border fault reached a maximum near the end of the Triassic, at which time South Mountain was the source of detritus deposited along the western edge of the basin in the form of extensive fanglomerate deposits (Stose, 1932, p. 92-93).

Since Triassic time the South Mountain area has been continuously above sea level. Denudation dominated, but sedimentation in the form of fluvial, lacustrine, or bog deposits occurred locally.

Terrestrial lignite, the only dated remnant of post-Triassic sedimentation, accumulated in a lake, pond, or fresh-water swamp during the Late Cretaceous (Pierce, 1965), and was subsequently covered by a thick gravel sequence (Figure 16, F). The gravels, deposited as alluvial fans which coalesced to form an alluvial apron along the margins of South Mountain, are post-Cretaceous and may relate mostly to Pleistocene sedimentation during periglacial conditions. Stream valleys incised into the terrace gravels were cut during Recent time by present-day streams debouching from South Mountain.

STRUCTURAL DYNAMICS AND MECHANICS

The structure of South Mountain is dominated by northeast-trending folds, and a southeasterly dipping flow cleavage and related downdip lineation. All of the rocks exposed in the study area have been deformed according to a uniform structural plan.

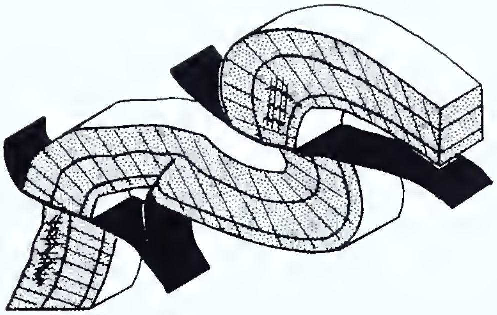
Deformation was largely the result of a stress field in which P_1 (maximum principal stress direction) was subhorizontal and oriented northwest-southeast. The structural plan reflects a period of tectonism in which the primary stress field maintained a constant orientation. The development of several structural elements aids in the recognition of different phases in the deformation plan.

The arcuate pattern of the Appalachians developed as early as the Early Silurian (Woodward, 1964, p. 340). The central Pennsylvania fold belt including South Mountain, symmetrical to a N40W axis which radiates from the area of the Baltimore dome, is interpreted as the result of compressional stress along this axis (Woodward, 1964, p. 341, fig. 3).

The initial phase (Figure 17) of South Mountain deformation included folding and faulting. Faulting was the result of northwest-oriented compression and northeast-southwest extension caused by the early development of an arcuate pattern in the folded belt. The strike of the Carbaugh-Marsh Creek fault, about N10W, and its left-lateral displacement is one potential shear direction under such a stress field. The difference in the intensity of the folding and faulting between the blocks north and south of the Carbaugh-Marsh Creek fault indicates that this structure

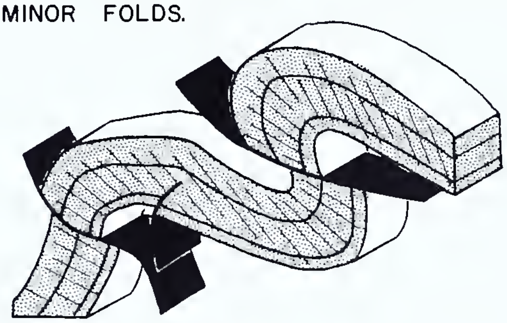
MAIN PHASES
SOUTH MOUNTAIN DEFORMATION
CALEDONIA PARK AREA

PHASE 5



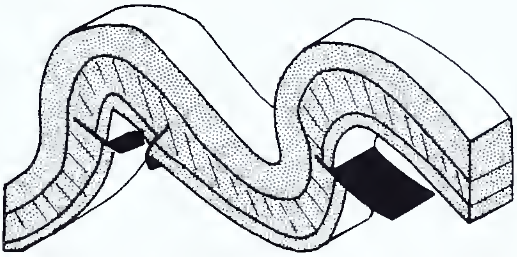
LATE FOLDING PHASE; LOCAL DEVELOPMENT OF SLIP CLEAVAGE (RIGHT) AND FRACTURE CLEAVAGE (LEFT) RELATED TO MINOR FOLDS.

PHASE 4



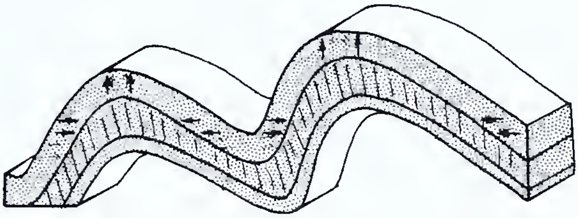
CONTINUED FOLDING AND FAULTING; SHEAR FOLDING BECOMES INCREASINGLY DOMINANT.

PHASE 3



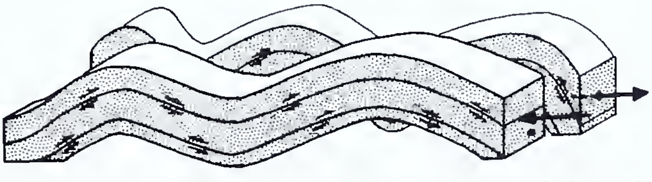
INITIATION OF FAULTING; EFFECT OF "ROOM-PROBLEM" IN CORE OF CONCENTRIC FOLDS AND STRETCHING OF FORELIMB IN OVERTURNED SIMILAR FOLDS.

PHASE 2



CONTINUED FOLDING; SHEAR FOLDING AND SUB-VERTICAL AXIAL-PLANE CLEAVAGE DEVELOP IN THE INCOMPETENT BEDS. FOLD GEOMETRY GOVERNED BY COMPETENT BEDS.

PHASE I



INITIAL FOLDING BY FLEXURAL-SLIP; INITIATION OF RIGHT-LATERAL DISPLACEMENT ON CARBAUGH-MARSH CREEK TRANSVERSE FAULT.

Figure 17. Structural development of the Caledonia Park area during Stage D of the geologic history.

developed early in the deformation plan. Left lateral displacement on the fault effected the rotation of structural elements and topography, and caused the truncation and termination of folds and longitudinal faults. The marked change in the trend of South Mountain near Caledonia Park is the only significant change in the structural and topographic grain of the Blue Ridge between northern Virginia and eastern Pennsylvania. It is likely that movement on this fault continued throughout most of the folding phase.

Both flexure folding and shear folding combined with laminar flow are recognized (Figure 17, phase 2). The dominance of one type or the other is a function of the competency, and the structural and stratigraphic position of the rock within the South Mountain anticlinorium. The size and shape of the folds were mainly determined by the total thickness of strata involved, and the thickness and rigidity of more competent members. Less competent beds followed the competent strata more or less passively, and were probably subjected to slightly different stress conditions.

It is likely that folds started as flexure folds, which involved bending of the strata and concentric shear. Differential slip on bedding, an integral part of the flexure mechanism (de Sitter, 1956, p. 182), resulted in the formation of slickensides on bedding surfaces oriented nearly perpendicular to the bedding fold axes. During this phase of folding, it is probable that the axial planes of the folds were nearly vertical or steeply east dipping.

At any stage in the development of an anticline by a concentric mechanism, there are two marginal blocks moving inward and a central block moving upward (de Sitter, 1956, p. 196-200). Compression of the less competent beds by the inward movement of the rigid marginal blocks led to adjustments in the form of flattening normal to the principal stress, and elongation upwards and normal to the main stress in the weaker rocks. Flattening was accompanied by recrystallization, distortion and alignment of oolites, quartz grains, and rock fragments, and differential slip on the cleavage surfaces. Although de Sitter (1956, p. 215) notes that differential movement on cleavage is not essential to the flattening process, slickensided, unfolded cleavage planes indicate that this type of movement occurred in the rocks of the Caledonia Park area.

The cleavage-forming mechanism resulted in the formation of a lineated flow cleavage parallel to the axial planes of similar folds. Limited slip occurred on bedding even after cleavage developed to produce a deflection of the cleavage near the margins of incompetent beds, however, the geometry of the folds is related primarily to flattening and shear on cleavage. The role of bedding in rocks effected by cleavage folding was passive.

With continued deformation, beds of increasing competence underwent shear or cleavage folding. The pervasive character of the flow cleavage (S_1) indicates that by the final stage of deformation all but the most competent rocks had been affected by folding mechanism.

The consistent orientation of *Skolithos* tubes perpendicular to bedding and the lack of any recognizable change in bed thickness attest to the concentric geometry of folds involving the Montalto and Antietam, the only units deformed entirely by a flexural slip mechanism.

Cleavage (S_1) does not occur in the Montalto and Antietam except at a few scattered localities. However on the steep western limb of the South Mountain anticlinorium, these rocks have partially responded to strain by developing joints having an orientation similar to that of the regional cleavage but at a much wider spacing.

Because of the considerable thickness and concentric geometry of folds developed in the upper Chilhowee sandstones and quartzites, a room problem developed in the core of these structures (Figure 17, phase 3). This crowding initially caused a broadening of the anticlinal crests, and is reflected in the broad Big Flat Ridge anticlinal structure (Plate 1). Further folding (Figure 17, phase 4) led to the upward "thrusting" of the crestal block along steeply inclined faults. The longitudinal faults near the perimeter of the Big Flat structure are interpreted as the result of this mechanism.

Other longitudinal faults such as that on the western edge of Piney Mountain (Plate 1) are likely of different origin. These faults are associated with the steepened or overturned limbs of shear folds. They are interpreted as break-thrusts, developed in response to a weakening of the forelimb by extension in the northwest-southeast direction.

Flow cleavage (S_1) in the incompetent beds was deformed during the later stages of deformation (Figure 17, phase 5). Initially deformation was by flexure with slip taking place on the flow cleavage planes, but with continued folding an axial plane cleavage (slip cleavage, S_2') formed in the axial region of the flow cleavage folds. The final form of the folds, chevron in transverse section, was the effect of combined slip on flow cleavage (S_1) and "micro-thrusting" along the axial plane slip cleavage (S_2') as shown by slickensides on flow cleavage planes normal to the fold axes (L_{1x2}') and the offset of the older cleavage (S_1) along the slip cleavage (S_2').

Bedding fold axes, cleavage fold axes, and the intersection of bedding with flow cleavage (L_{0x1}) and flow cleavage with slip cleavage (L_{1x2}') are all parallel, a result of a stress field of constant orientation during the development of bedding and cleavage folds. In this primary field, P_1 was oriented northwest-southeast, P_2, northeast-southwest, and P_3 vertically.

As the major structures developed, secondary stress fields of new orientation were established. Longitudinal extension induced by the plunge and arcuate plan of South Mountain in south-central Pennsylvania prevailed locally, at times, over the waning regional stress system during the final phases of deformation.

Related to these local secondary stress fields, in which P_3 was horizontal, are some flank faults located principally along the margin of the Big Flat Ridge anticlinal structure. These faults are typically steep-dipping and show offset caused by strike-slip or normal faulting. A secondary stress field with P_1 vertical (lithostatic pressure) would cause gravity faults, but strike-slip faults would result if P_1 were horizontal.

Nearly vertical conjugate joint sets occur in the area. The orientation of some of these joints can best be explained as conjugate shear planes related to a stress field in which P_3 was horizontal and oriented northeast-southwest, and P_2 vertical.

In the overturned or steep western limb of the South Mountain anticlinorium, carbonate beds were locally folded about a near-horizontal axis to form recumbent folds with a sub-horizontal axial plane. The more competent were folded by flexural slip; the incompetent by shear along cleavage (S_2). These structures may be explained in terms of local gravitational slumping under the influence of a sub-vertically directed principal stress caused by the increased overburden as the western flank of the anticlinorium continued to steepen.

To what extent was basement involved in the South Mountain deformation plan? Divergence in orientation between the gneissic banding of the Precambrian granitic rocks and the flow cleavage (S_1) is evidence for a pre-Catoctin period of deformation. However, southeast dipping flow cleavage (S_1) and downdip lineation (L_a) are present in the basement rocks of Maryland and Virginia (Cloos, 1957, p. 835-836; Nickelsen, 1956, p. 242; Reed, 1955, p. 877). Thus the occurence of the characteristic "South Mountain" cleavage and lineation in all these rocks is strong evidence of basement involvement during the Paleozoic orogenic event. If an important structural detachment (décollement) occurred during the development of the South Mountain fold, it is not likely to be located at the basement-Catoctin or Catoctin-Chilhowee contact.

The structural elements which typify South Mountain maintain a consistent character and orientation through the Chilhowee Group and the overlying Cambro-Ordovician carbonates into rocks of Beekmantown age. The recumbent folds and related sub-horizontal cleavage in some of the carbonate beds are the only structures not duplicated in the underlying sequence. These local structures do not represent a significant change in the structural style across the Blue Ridge - Valley and Ridge boundary. It is evident that the rocks in the Caledonia Park area have all participated in a uniform deformation plan.

SUMMARY AND CONCLUSIONS

This study has provided much additional data on South Mountain in Pennsylvania. Analysis and evaluation of these data have led to an interpretation of the nature, extent, and sequence of Paleozoic events including structural genesis and tectonic evolution, in harmony with the regional tectonic design.

Subdivision of the Catoctin, Loudoun and Harpers Formations into smaller lithologic units than previously recognized is based upon mapping. These lithologic subdivisions have been mapped over an area which extends about 15 miles parallel to the regional strike. Utilization of these informal divisions has outlined structural details not otherwise apparent, and led to revised structural interpretations for several localities previously studied by Stose (1909, 1929, 1932).

The "blue quartzite" of the Montalto Member (Harpers Formation) is a key bed which has been identified by the writer at the same stratigraphic position throughout much of the study area, and at other localities for a distance of 30 miles parallel to the South Mountain axis. The presence of this unit makes it possible to differentiate between the lithologically similar quartzites of the Montalto and Antietam not only in the area of this report, but possibly also in the unfossiliferous portions of the Cambrian section elsewhere in south-central Pennsylvania.

Concentric and similar folds occur in the area. Folds in the Montalto and Antietam are concentric and developed by flexural slip, but the other strata have been folded by a combination of the flexural slip and shear mechanism; the dominant mechanism is a function mainly of structural competence.

Folding was probably initiated in all rocks by flexural slip along bedding but graded into cleavage (shear) folding during later stages of deformation in rocks less competent than well-cemented protoquartzites and quartzites. Although folding was dominated by shear, active bedding plane slip occurred locally in the less competent rocks to produce offset and deflection of the axial plane or flow cleavage (S_1) at bed boundaries.

The intensive development of flow cleavage (S_1) obliterated the original stratification of the more incompetent rocks and terminated flexure by slip on bedding. In the phyllites, flow cleavage was eventually deformed by flexural slip along cleavage surfaces. A second cleavage, slip cleavage (S_2'), developed in response to continued folding of the older cleavage. The combination of active flexural slip on flow cleavage (S_1) and shear along slip cleavage (S_2') completed the deformation of the older cleavage (S_1) and generated small-scale chevron folds.

The axial plane cleavage (S_1) and related cleavage or shear folds are the products of plastic deformation related to a single stress field with P_1 oriented northwest-southeast.

Peripheral granulation and the development of crushed quartz "tails" resulted in detrital quartz grains becoming shortened normal to and lengthened down the dip of the axial planes of the folds and the flow cleavage (S_1). In the metarhyolites micropoikilitic patches attenuated by plastic flow, and feldspar phenocrysts elongated through the development of tension fractures indicate a similar adjustment and alignment. These structures, analogous to the deformed oolites described by Cloos (1947) and also found in the area of this report, caused a distortion of rock thickness in all but the most competent rocks of the area.

Extensive quartz grain deformation is not restricted to rocks involved in shear folding. Extreme undulose extinction, deformation lamallae, peripheral granulation, and recrystallization of detrital quartz grains is also prominent in rocks concentrically folded.

A series of northeast-trending folds superimposed upon the western limb of the South Mountain anticlinorium is the internal structure of South Mountain in Pennsylvania. Folds are asymmetric or overturned to the west, and exhibit an axial plane cleavage which dips southeast. Bedding-flow cleavage relations and cross-bedding reveal more overturned sections than previously recognized, particularly in the Weverton Quartzite.

Although folding is the dominant mode of deformation, faulting is common and occurred during all phases of the deformation.

The larger faults are generally steep-dipping, longitudinal structures genetically related to "crowding" in the core of concentrically folded anticlines, and to the stretching and overturning of fold limbs in shear folds. These dislocations developed during the main folding phase and post-date the inception of the Carbaugh-Marsh Creek fault. This structure, a right lateral transverse fault, caused folding of dissimilar intensity to develop in the adjacent blocks, produced an early displacement of structural axes and cleavage (S_1) and effected a change in the trend of South Mountain near Caledonia Park.

Longitudinal stretching related to the axial plunge of anticlinal structures and the arcuate plan of South Mountain set up secondary stress conditions (P_3, northeast-southwest) which caused small-scale oblique faults on the flanks of anticlines.

Local recumbent concentric and shear folds developed in the carbonate terrane are interpreted as gravity collapse structures. These folds formed near the close of the regional deformation, certainly after the characteristic South Mountain deformation plan had been impressed upon all rocks in the area.

Fold axes and Pi diagrams of bedding define a regional plunge southwestward. This is in harmony with the fact that this segment of the Blue Ridge is located just north of the deepest point of an axial depression of the anticlinorium between culminations near Mount Holly Springs, Pennsylvania, to the north, and Harpers Ferry, West Virginia to the south.

The mapped area is an integral part of the Blue Ridge structural province, a belt north of the Potomac River characterized by a uniform deformation style (Cloos, 1947, 1957). This plan is dominated by a regional northeast trending flow cleavage (S_1) and related southeast striking downdip lineation whose development and orientation is uniform in rocks of diverse composition throughout this region. However some variations in the character of the deformation occur in the Caledonia Park area compared to more southern segments of this belt.

(1) Flow cleavage and the downdip lineation are inclined more steeply southeast, a continuation of the progressive steepening of the axial planes of folds north of Harpers Ferry, West Virginia.

(2) Slip cleavage (S_2') is more restricted stratigraphically and occurs only in phyllites of the Loudoun and Catoctin Formations. Similarly, the regional flow cleavage (S_1) is only locally developed in the Montalto and Antietam.

(3) Concentric geometry, developed through flexural slip along bedding, characterize the folds developed in the most competent clastic sediments.

(4) Faults are more extensively developed, particularly those oriented transverse to fold axes.

These variations signify less intense deformation of a stratigraphic section of more "brittle" character, interpreted to be, in part, the result of the northward increase in the competence of the stratigraphic section as the graywacke siltstone of the Harpers Formation is gradually replaced by protoquartzites and quartzites of the Montalto Member, and the quartz content and thickness of the Antietam are increased. An additional factor may have been a decrease in the deformative stress in magnitude or time.

Uniform development and orientation of structural elements in all stratigraphic units in the Caledonia Park area indicate the lack of any important structural discontinuities or décollement in this segment of the Blue Ridge. The occurrence of the regional cleavage and lineation in basement rocks to the south supports basement participation in the deformation.

Parallelism of structural axes of different age describes a stress field of constant orientation. Tentative correlation of flow (S_1) and slip (S_2') cleavage in the Caledonia Park area with similar structures associated with dated rocks in the Piedmont Province east of South Mountain suggests that deformation may have been a Late Devonian event. Evidence from rocks in the study area, however, can define the orogenic period no more closely than post-Beekmantown to Triassic. Hence it is evident that Ordovician (Taconian) or Late Paleozoic (Appalachian) events may be responsible for the South Mountain deformation, or at least represent phases in the deformation plan.

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APPENDIX A — PETROGRAPHY

PETROGRAPHY OF THE CATOCTIN FORMATION

Metabasalt

In thin section, the metabasalt is holocrystalline. Amphibole, epidote, chlorite, hematite, leucoxene, sphene, plus small amounts of secondary quartz usually fill the interstices between laths of plagioclase which comprise the groundmass. Some laths outline a felty or less commonly, a pilotaxitic fabric (Plate 5, Fig. 5).

Groundmass plagioclase is primarily subhedral, and ranges in size from 0.05 - 0.75 mm. The laths of plagioclase are usually between 0.05 - 0.25 mm long, but average 0.2 - 0.5 mm in the amygdaloidal rock. Albite twinning is the most common, but it is often combined with Carlsbad twinning. Some laths are partially altered to micromicas, chlorite, and quartz.

The plagioclase has been identified as albite by 2V measurements made on the universal stage, and by extinction angles on the albite twins as determined on the flat stage. Thus the plagioclase determination is in agreement with Nickelsen (1955, p. 245) and Reed (1955, p. 889), but conflicts with that of Stose and Bascom (1929, p. 5), and Bloomer and Bloomer (1947, p. 48) who identify the plagioclase as andesine.

The amphibole in the metabasalt is probably actinolite. It generally occurs as fibrous, sub-equant to prismatic grains, 0.15 - 0.40 mm long. Moderately low birefringence (0.012-0.016), colorless to very pale-green pleochroic colors, and extinction inclined 12-20 degrees to the cleavage trace are its chief characteristics.

Large (0.5 - 2.0 mm), equant, poikiloblastic grains of amphibole are found to correspond to the "green spots" in the coarse-grained metabasalt. Epidote, hematite, leucoxene, and sphene are encased by the amphibole, a major portion of which is altered to chlorite.

A dimensional alignment of the amphibole is observed in some specimens. This orientation is nearly parallel to that of the groundmass plagioclase when both are present. As the mineral orientation is nearly parallel to the hand-specimen foliation, the alignment is evidently a reflection of the pervasive flow cleavage which is characteristic of the area.

Epidote is almost ubiquitous, occurring either as small (0.15 - 0.30 mm), subhedral to anhedral grains, or in clusters of three to seven grains. The clusters have an average diameter of 0.5 - 0.8 mm.

Much of the chlorite occurs as minute prisms or flakes in the groundmass, or as interstitial patches. In the coarse-grained metabasalt, however, it occurs as an alteration product of the amphibole, replacing as much as ⅓ - ⅔ of the amphibole crystals. The remaining quantitatively important minerals include hematite, leucoxene, and sphene. These minerals are usually equigranular, and occupy an intergranular position relative to the albite laths of the groundmass. Hematite commonly encloses grains of epidote, imparting a brilliant red stain along the periphery which may be mistaken as piedmontite. Together, these minerals may constitute more than 15-20 percent of the rock.

The porphyritic metabasalt contains relatively fresh phenocrysts of albite, intergranular hematite, leucoxene, sphene, and chlorite set in a mat of interlocked laths of plagioclase. Epidote and amphibole are conspicuous by their almost complete absence. Isolated patches of this rock carry hematite, leucoxene, and sphene in excess of 15-20 percent.

The phenocrysts are typically euhedral to subhedral. Pericline twinning occurs together with the more common albite and Carlsbad twinning (Plate 5, Fig. 6). Many of the albite phenocrysts are strained as shown by irregular extinction of the twin lamellae, incipient fracturing, and bending of the crystal. In addition, longitudinal zones or patches of low birefringent, cryptocrystalline to microcrystalline material are common to many of the feldspars (Plate 5, Fig. 7). These zones may represent glass inclusion or be an exsolution phenomenon. A diffuse optic figure obtained on a few of the largest particles suggests a biaxial character.

Amygdules, the only definite primary structure recognized in the metabasalts of the area, are normally filled by albite or epidote, but some show a filling of albite in combination with a center of vermicular chlorite or quartz (Plate 5, Fig. 8).

Both the albite and epidote in the amygdules are subhedral. They occur as smaller, less well-developed grains at the periphery of the amygdule, but become larger and more euhedral radially inward. The feldspar laths are characteristically unaltered and water-clear.

The amygdaloidal rock is mainly composed of partially altered, fibrous amphibole, and interstitial hematite, leucoxene, and sphene in a groundmass of plagioclase. Secondary epidote and quartz are usually present in small quantities.

The groundmass plagioclase of the amygdaloidal variety is generally unaltered, and is decidedly coarser than that of the non-amygdaloidal rocks.

The mineral composition of some of the metabasalts is given in Table 3. Percentages determined by point counting are probably only accurate to within a few percent owing to the general fine-grained character of the rocks. For comparative purposes, similar data condensed from Reed (1955, p. 889) for the Luray area, Virginia, where the deformation of the metabasalts is less, is also included in Table 3.

Table 4 is a tabulation of those chemical analyses of the metabasalt available in the literature.

Epidosite

In thin section, epidote (40-60 percent), quartz (25-40 percent), and leucoxene and sphene (7-12 percent) prove to be the essential minerals. Amphibole and specular hematite occur in small quantities (<5 percent).

The rock is holocrystalline. Clusters of subhedral epidote, about 1.4 mm in diameter, are located in a variegated groundmass of quartz and epidote. The quartz occurs as microcrystalline aggregates or as interlocked, undulose grains 0.5-0.6 mm in diameter. Greenish, microcrystalline patches of epidote, and scattered grains of amphibole and hematite, usuallly associated with areas of quartz, complete the groundmass.

In the amygdaloid varieties, the vesicles are surrounded by an external zone of concentrated epidote, leucoxene, and sphene. Internally, subhedral epidote surrounds a core of clear, medium-grained quartz.

Amphibole and hematite are scattered throughout the groundmass, usually in association with areas of quartz.

TABLE 3. Point Count Data - Catoctin Metabasalt (in percent)

No.	Albite	Chlorite	Epidote	Actinolite	Pyroxene	Iron Oxides	Quartz	Sphene & Leucoxene
1	34	12	14	11		7	5	17
2	38	19	3	14		13	2	9
3	67	3				12	1	16
4	31	16	18	17		10	1	7
5	35	18	12	9	6	9	1	8
6	36	12	14	15	5	13	1	6
7	16	27	8	16	14	6	1	10

1 Even-grained metabasalt, 1.0 mile northwest of Strasbaugh School
2 Amygdaloidal metabasalt, west side Pa. 234, 0.6 mile northeast of Strasbaugh School
3 Porphyritic metabasalt, float, elevation 1360′ along dirt road 0.7 miles southeast of Strasbaugh School
4 "Spotted" metabasalt, east side Rocky Mountain Creek, 2.0 miles south of Caledonia Park
5 Porphyritic greenstones (avg. of 6); Luray area, Virginia
6 Coarse-grained greenstones (avg. of 2); Luray area, Virginia
7 Fine-grained greenstones (avg. of 3); Luray area, Virginia

TABLE 4. Chemical Analyses of Catoctin Metabasalt from South Mountain, Pennsylvania

	1	2	3	4	5	6
SiO_2	51.89	50.36	46.79	44.46	49.67	48.57
Al_2O_3	15.28	17.37	14.22	13.93	13.65	14.89
TiO_2	0.91	3.10	1.66	3.97	1.40	1.61
Fe_2O_3	3.10	8.13	5.10	6.80	8.32	6.29
FeO	3.60	4.33	9.42	8.59	5.19	6.23
MnO	0.12	0.24	0.18	0.28	0.16	0.20
MgO	8.68	3.57	5.86	5.82	6.37	6.06
CaO	7.38	4.29	10.14	6.85	10.39	7.81
Na_2O	3.27	4.74	2.38	2.89	1.11	2.88
K_2O	2.57	0.08	0.77	1.48	0.04	0.99
H_2O+	1.17	2.73	2.98	3.43	7.41	3.54
H_2O^-	1.37	0.16	0.54	0.04	0.00	0.42
P_2O_5	0.61	0.63	0.35	1.01	0.12	0.54
Total	99.95	99.73	100.39	99.65	99.83	99.89

1 Catoctin schist. Bechtel Copper Shaft, South Mountain, Pa. (Stose and Stose, 1946, p. 21, quoted from G. H. Williams)
2 Metabasalt from northwest of Devil's Race Course, Gum Spring Road, Fairfield Quad., Pa. (Stose, 1932, p. 40)
3 Metabasalt from 2½ mi. west of triangulation station on Green Ridge, Fairfield, Quad., Pa. (Stose, 1932, p. 40)
4 Metabasalt, Iron Springs 7½′ Quad., South Mountain, Pa.
5 (Landy, R. A., 1961, p. 93)
6 Average 1-5

Bluish Metarhyolite

Subhedral feldspar, occurring individually or in clusters of a few laths, is the most common phenocryst in the metarhyolites. Although Bascom (1896, p. 40) has identified the feldspar phenocrysts as anorthoclase, orthoclase, and microperthite, the writer identifies only albite. This identification is based on the positive optical character, negative index, and large optic angle (75-85) of the mineral.

Probably the most distinctive characteristic of the albite is its chess-board twin structure. "Chess-board albite is distinguished by a characteristic development of the albite law resulting in a complex pattern of short alternating twin lamellae which, unlike usual twin lamellae, do not pass through the entire crystal. Consequently when seen in thin section, cut perpendicular to (010), under crossed nicols, a chess-board pattern results when one set of twins is extinguished." (Starkey, 1959, p. 141) An example of this structure is shown in Plate 6, Fig. 1. Chess-board twinning, which probably represents the microperthite of Bascom, may be observed in combination with Carlsbad twinning.

The feldspars commonly contain inclusions of very fine-grained quartz, hematite, and other opaques. Those laths which have an orange color in hand specimen tend to have a cloudy, tan appearance in thin section owing to the inclusion of extremely fine hematite. Many of the laths show fractures and micro-faulting (Plate 6, Fig. 2). Where a phenocryst has been pulled apart, the open fracture is filled with quartz (generally coarser than that of the groundmass) and sericite. Normally the feldspar shows some alteration to clays and micromicas.

The quartz phenocrysts are round to almond-shaped. Undulose extinction, fluid and mineral inclusions, and moderate fracturing characterize the quartz. Where kinetic metamorphism has been most effective, the phenocrysts are highly strained and fractured, and may show a "tail" of finely divided quartz at one end. The penetration of the borders of the quartz by the groundmass occurs in nearly all cases. However, this phenomenon is more pronounced in samples showing intense physical disruption. There the quartz phenocrysts are punctured by patches of groundmass, and in some cases, almost totally replaced by the groundmass.

The groundmass mineralogy of the metarhyolite is relatively simple. Quartz, plagioclase, hematite, and muscovite (including sericite) are the only essential minerals. A very fine-grained, microgranitic aggregate of quartz and feldspar, frequently modified by the occurrence of fluidal, micropoikilitic, or spherulitic structures, is typical of the rock. Bascom (1896, p. 47) defines the micropoikilitic structure as "irregular quartz patches, inclosing microlites of lath-shaped feldspars or other minerals of independent orientation, which gives a pronounced mottled or "patchy" appearance to the groundmass . . . ." These patches are usually outlined by finely granular hematite (Plate 6, Fig. 3). If sericite is developed, this mineral frequently wraps around the micropoikilitic areas. The basic shape of these patches is equant, but some rocks show only elongated patches. The average length of these patches is about 0.5 - 1.0 mm.

A few relatively coarse grains of feldspar in the groundmass have been identified as albite. These grains, like the phenocrysts, are commonly bordered by sericite. Sericite also occurs associated with hematite in irregular patches in the groundmass, and as part of the matrix in the brecciated metarhyolites.

Spherulitic structures consist of fan-shaped aggregates which tend to be arranged in bands or zones, but may occur singly. The size of an individual spherulite may reach 0.3 mm, although the average is probably only half this size. Owing to the small size of the individual spherulite components, positive identification of the mineral constituents is difficult. This structure occurs sparingly in the bluish metarhyolites.

The banding in the metarhyolites, mainly due to the concentration of the very fine hematite granules, is also produced by quartz-rich zones, and alternating quartz-feldspar bands of variable sericite content. Where the banding is related to quartz content, the quartz grains are coarser than those of the groundmass.

In the vesicular rock, the vesicles, filled by small, interlocked quartz grains, commonly contain a centrally located grain of sphene(?). A rind of very fine, granular hematite encloses the amygdules which range in size from 0.15 - 1.5 mm.

Specular hematite is one of the most abundant minerals in the metarhyolites. It occurs as fine to extremely fine granules as well as anhedral grains up to 0.5 mm long. Very rarely does the hematite content of these rocks fall below 8-10 percent, and 15 percent hematite content is not uncommon.

Other mineral components of the rocks are the mineral pair leucoxene-sphene, plus chlorite, biotite, epidote, and zircon. Except for the rocks in which piedmontite occurs in veins and seams, and in the brecciated rock, this mineral is also an accessory. In the breccia where it combines with quartz as the cementing material, the piedmontite occurs as vividly pleochroic, subhedral, prismatic crystals and columnar aggregates up to 0.75 mm long (Plate 6, Fig. 4). The amount of piedmontite in this rock is estimated to be 15-18 percent.

Mottled Metarhyolite

Microscopic examination shows the essential minerals to be quartz, plagioclase, and hematite. Varietal minerals include ilmenite, sphene, sericite, and, in the rock lineated in *a*, calcite. Zircon, biotite, and epidote are accessories.

Generally equant, micropoikilitic quartz-feldspar patches with a diameter of about 0.5 - 0.8 mm comprise the holocrystalline groundmass. The patches often have faint, partial rims of finely granular hematite. The groundmass minerals show little visible alteration.

Phenocrysts of quartz and feldspar comprise as much as 10 percent of the rock. The round to almond-shaped quartz phenocrysts are characterized by fine, arcuate fractures, abundant fluid and microlite inclusions, and mild undulose extinction. Many of these grains have indistinct borders and are marginally replaced by minute penetrations of groundmass material.

Quartz also occurs as irregular grains (0.5 mm diameter) in sinuous trains in the groundmass.

The phenocrysts of feldspar are identified as albite on the basis of optical properties. The laths are subhedral and usually show little alteration except for fine, marginal embayments filled by the groundmass. Chess-board structure, alone or combined with Carlsbad twinning, is very common.

Albite occurs as single laths or, less commonly, in groups of three to five grains. Many of the feldspar grains contain inclusions of independently oriented feldspar laths, hematite, and very fine-grained quartz.

Thin sections from the "lineated" rock are characterized by elongate, micropoikilitic patches whose length of 1.0 - 4.5 mm is many times the width. The direction of groundmass elongation is closely related to the *a* lineation observed both in hand specimen and outcrop.

Other textural differences can be distinguished in the "lineated" rock. The micropoikilitic areas are enclosed by prominent rims of hematite and are commonly separated along each side by thin, discontinuous streams of sericite. Quartz phenocrysts are highly undulose and contain many hematite-filled fractures (Plate 6, Fig. 5). Some of the quartz grains show "tails" of fine, crushed quartz aligned parallel to the major axis of the elongate groundmass patches.

Fractures in the feldspar phenocrysts of the "lineated" rock are also more highly developed. Open fractures are filled by quartz, hematite, or sericite. Filled fractures oriented normal to the groundmass lineation are interpreted as extension fractures. Calcite, as irregular patches within the feldspar laths and as a marginal embayment filling in the quartz phenocrysts, appears only in the "lineated" rock.

Parallelism between phenocrysts orientation and groundmass elongation is not strong. However, where the quartz and feldspar are positioned oblique to the lineation, the groundmass shows local disruption (Plate 6, Fig. 6).

Metarhyolite Breccia

The phenocrysts in this rock are generally altered to sericite and piedmontite. Chess-board structure, in some grains combined with Carlsbad twinning, can be recognized in the less altered subhedral laths. The feldspar is optically identified as a plagioclase, probably albite.

A microcrystalline groundmass, with either a micropoikilitic or microgranitic texture, is composed chiefly of quartz and feldspar. Small (<0.15 mm), anhedral grains of hematite, leucoxene, sphene, and quartz, plus patches of sericite, and subhedral piedmontite complete the mineralogy of the fragments. Only quartz, feldspar, and hematite are essential minerals.

Hematite (<50 percent), piedmontite (10-15 percent) and sericite (10-12 percent), plus smaller quantities of plagioclase, quartz, leucoxene, and finely crushed rhyolitic material comprise the matrix. The most distinctive mineral in the matrix is piedmontite. It displays brilliant gold, red-orange, and crimson pleochroic colors, and occurs as grains up to 0.5 mm in length.

Red Porphyritic Metarhyolite

The feldspar phenocrysts, when examined microscopically, are typically equant to lath-shaped, and have a subhedral to anhedral form. All exhibit chess-board structure; some may also show Carlsbad twinning. The optical character of the phenocrysts is that of albite.

Fine cracks filled by hematitic dust are well developed in the feldspar. In addition, inclusions of very fine-grained, low birefringent material (quartz?) and peripheral indentations filled by the groundmass material mark many of the phenocrysts.

Quartz phenocrysts have a round to elliptical habit, and show numerous, distinct fractures which cross the grain in broadly curving arcs. These cracks are filled wtih very fine hematite which probably gives the quartz its rich red color. Quartz grains exhibit moderate to intense undulose extinction and many are embayed along the edges by the groundmass, with rare grains showing significant replacement by groundmass material.

Phenocrysts comprise an estimated 10 percent of the rock, with feldspar content two or three times greater than quartz.

The groundmass is holocrystalline. Micropoikilitic patches, composed chiefly of intergrown quartz and feldspar, constitute nearly 80 percent of the matrix. These patches have a characteristic shape in each rock; either equant or elongate. Outside the micropoikilitic areas, quartz, hematite, and sericite are the most abundant minerals. Quartz occurs as sparse, small irregular grains in the groundmass. Sericite usually is found as a discontinuous film around the edges of the micropoikilitic patches, as a filling of open fractures in feldspar laths, and associated with quartz in patches less than 1.0 mm in diameter. The sericite content is usually less than 3 percent except in sections showing an elongate matrix, where it exceeds 5 percent.

Hematite constitutes at least 10-15 percent of the rock. It is of two types: very-fine granules filling open fractures in feldspar and quartz and forming a nearly continuous fringe around the patches of the groundmass, and longer, anhedral grains up to 0.4 mm. long.

Much of the finer grained hematite is associated with very fine-grained sphene and leucoxene. This mineral pair usually does not exceed 5 percent. Zircon, epidote, and piedmontite are accessory minerals.

Phyllites

The phyllites are very fine-grained aggregates of aligned, slightly pleochroic, colorless to pale green shreds of sericite (pyrophyllite?), muscovite, and quartz. The diameter of the quartz does not exceed 0.1 mm even in the coarsest phyllites.

Leucoxene and sphene accompany quartz and sericite (pyrophyllite?) in all thin sections studied, and probably constitute about 1 to 3 percent of the rock. Other minerals identified include chlorite, feldspar, zircon, epidote, and hematite. These minerals are accessory minerals in every case but one, although they are not found in every sample. Epidote occurs as subhedral, equant grains about 0.15 mm in diameter in one sample, and comprises about 3 percent of the rock.

PETROGRAPHY OF THE CHILHOWEE GROUP

Loudoun Formation

Fine to microcrystalline quartz, muscovite, sericite, chlorite, and hematite are the chief constituents of the phyllites.

Many of these rocks contain fine-grained rock fragments and detrital quartz. The detrital quartz is inequant, subangular to subrounded, and poorly sorted. Diameter size ranges from 0.2 - 3.0 mm. Intense undulose extinction, deformation lamellae, and peripheral granulation are developed in most of these grains. Some of the larger quartz grains exhibit fracture patterns and a clearness which resembles the quartz phenocrysts in some of the metarhyolites — a potential source for this mineral.

The matrix of these phyllites is composed of a well-oriented, fine-grained aggregate of quartz, muscovite, and sericite. Chlorite is locally abundant, occurring mainly in scattered zones in which the chlorite laths form a mat-like aggregate (Plate 6, Fig. 7).

Elongate blebs and streaks observed in hand specimens are fine-grained, muscovite-chlorite-quartz mosaics with a discontinuous rim of chlorite. Small, subhedral grains of epidote occur within some of these structures, all of which show a marked parallelism with the groundmass.

Hematite is an essential constituent of all phyllites examined. It occurs as anhedral grains 0.02 - 0.5 mm in diameter scattered uniformly throughout the rock or concentrated in irregular bands. Leucoxene and hematite together may comprise 20 percent of the sample. Zircon, epidote, and subhedral to euhedral apatite are the common accessories. Pyrite is abundant in a gray phyllite which occurs along the lower eastern slopes of East Big Flat Ridge.

Thin sections of the conglomeratic beds show pebbles of quartz, quartzite, rhyolite, and phyllite plus smaller grains of quartz in an aligned matrix of quartz, muscovite, and sericite (Plate 6, Fig. 8).

Only the phyllite fragments exhibit clear downdip orientation in the cleavage plane. Most of the rhyolite pebbles are moderately altered; usually feldspar phenocrysts in these pebbles have been converted to micromicas and clay.

Quartz grains show poor sorting. The grains are subrounded to subangular, subequant to equant, and exhibit moderate to intense undulose extinction, granulated borders, and some deformation lamellae. Kink bands developed in some of the muscovite flakes indicate the physical deformation of this mineral.

Hematite occurs in anhedral grains up to 5 mm in diameter, and is an essential constituent of the conglomerate. Leucoxene is a common varietal mineral, whereas tourmaline, sphene, epidote, biotite, and zircon are accessories.

Point count data are given in Table 5 for five samples from this formation.

TABLE 5. Point Count Data — Loudoun Formation
(in percent)

Sample	Quartz	Mica	Chlorite	Rock Fragments	Heavy Minerals	Crushed Quartz	Micromicas
1	16	18	8	8	18		32
2	38	20	3	5	26		8
3	36	1		13	12	15	23
4	18	1		28	16	11	26
5	33			21	7	19	20

1 Phyllite member, about 0.7 mile southwest junction Piney Mountain Ridge Road and Miltonburger Road.

2 Phyllite member, about 0.4 mile west intersection Route 233 and Shippensburg Road.

3 Conglomerate member, southeast slope Piney Mountain about 0.8 mile west of Strasbaugh School.

4 Conglomerate member, elevation 1420′ southeast slope of Piney Mountain at power line 0.5 mile north of Shippensburg Road.

5 Conglomerate member, outcrop southeast slope East Big Flat Ridge 0.5 mile west of intersection of Route 233 and Shippensburg Road.

Weverton Quartzite

The suite of thin sections studied mostly represent Weverton samples from exposures along East Big Flat Ridge and Piney Mountain (Plate 1).

Lithologies in the lower three-fourths of the Weverton Quartzite include phyllitic sandstone and non-phyllitic sandstone plus subordinate conglomerates. For the most part the rocks may be classified as quartzose graywackes, protoquartzites, and quartzites.

In general, the phyllitic sandstones are graywackes composed of quartz grains (40-65 percent), rock fragments (1-10 percent), muscovite (0-8 percent), feldspar (0-3 percent), and matrix (25-45 percent). Size sorting is fair, and graded bedding is not uncommon.

Quartz occurs as subangular to subrounded, equant to subequant grains, 0.2 - 2.0 mm in diameter. In nearly every section, the quartz exhibits the effects of strain and cataclasis. Grains are characterized by moderate to intense undulose or patchy extinction. Deformation lamellae can be observed in most thin sections of the phyllitic sandstones, though less so than in other parts of the Weverton.

Margins of the quartz commonly show evidence of granulation resulting in the development of lenses or "tails" of finely comminuted quartz. These "tails" extend behind the parent grains and are parallel to the oriented minerals of the matrix. Additionally, the periphery of many quartz grains show some replacement by the matrix. Where the majority of the quartz is subequant, a recognizable dimensional orientation occurs (Plate 7, Figs. 1, 2).

Quartzite, volcanic (rhyolite?), shale-phyllite, and microcrystalline quartz (chert) fragments are recognized in the Weverton samples studied. A few thin sections containing highly altered, granular mosaics, in which quartz is the only identifiable mineral, may represent weathered granite fragments.

Generally only quartzite and volcanic fragments comprise more than 2 percent of the rock, and normally quartzite exceeds 50 percent of the total rock fragment content. Volcanic and shale-phyllite particles are more abundant in the lower part of the section whereas chert is more prominent near the top.

Less than half the samples contain muscovite and feldspar. The muscovite is generally subequant to elongate and is commonly wrapped about or wedged between grains of quartz.

Microcline is the main feldspar; albite has been identified in only a few samples. Both types occur as equant laths, and although some grains are fresh, most have suffered moderate alteration. Alteration is particularly concentrated along the margins and cleavage traces of the grains. Rarely does the feldspar content exceed a few percent.

The matrix is usually well aligned and is composed of crushed quartz, muscovite, sericite, and more rarely chlorite. The quartz is mainly derived through the granulation of the margins of the quartz grains. Shredded muscovite plus sericite probably result from the reconstitution of an original argillaceous matrix. Usually the micaceous minerals form slightly more than half of the matrix, although the crushed quartz may be more abundant locally.

The non-phyllitic sandstones differ from the phyllitic sandstones described above in a few significant respects. These rocks, mainly quartzose graywackes and protoquartzites, are generally well-sorted and have a quartz content of 75-85 percent. The quartz grains are characterized by severe strain, including a pronounced development of deformation lamellae and undulose extinction. In addition, some quartz grains are ruptured and the fractures filled by crushed quartz.

Quartzite fragments in small quantities occur in all samples, but volcanic and micaceous rock fragments are very rare. Chert is also commonly present but in very limited quantities.

A small percentage of the rock is matrix (5-15 percent). The lower matrix to grain ratio restricts the matrix to patches filling the interstices between grains. Although the mineral composition of the phyllitic and non-phyllitic rocks is similar, a good alignment of the matrix is not developed in the non-phyllitic rocks.

The sandstones of the upper member are dominated by quartz (70 - 95 percent) but variable quantities of feldspar, rock fragments, and muscovite also occur. These rocks are mainly protoquartzites and quartzites.

Quartz ranges in size from 0.25 - 8.0 mm in diameter with size sorting generally good. These grains are strongly undulose, fractured, and commonly contain well-formed deformation lamellae. Quartz grains in contact may show sutured margins.

Plagioclase (albite) and microcline are both present. Microcline is dominant however, and is found as small, anhedral, equant grains only a few of which have been highly altered. Probably the altered feldspar represents the clayey, rectangular bodies observed in hand-specimen and outcrop.

Rock fragments may constitute as much as 10 percent of the rock. At least 80 percent of the fragments are quartzite; the remainder is chert and very minor amounts of micaceous and volcanic particles.

The matrix occurs as intergranular patches or thin films around the grains. The matrix minerals are crushed quartz, sericite, and rarely chlorite. In places the sericite has penetrated and replaced the margins of the quartz grains.

The mineral composition of a representative group of thin sections from the Weverton is given in Table 6.

Harpers Formation

Thin sections representing all portions of the Harpers Formation were examined. Most of these sections were from samples collected on Piney Mountain, Green Ridge, Big Pine Flat, and Stony Knob (Plate 1).

The great majority of the rocks below the upper Harpers are texturally and compositionally quartzite as defined by Krynine (1948, p. 137, 149-152) whereas rocks in the upper Harpers are primarily low-rank graywackes as defined by Krynine (1948, p. 137, 149-153).

The rocks of the lower Montalto are composed almost wholly of detrital quartz (80-95 percent). Other minerals occurring as grains include very minor amounts of muscovite and rock fragments. Highly altered rectangular grains observed in hand samples, and thought to be weathered feldspar could not be recognized in the thin sections studied.

The quartz grains are normally subrounded and equant, and range in size from about 0.15 - 2.0 mm. Size sorting is estimated as fair. Most grains have moderately to intensely developed undulose extinction, and some grains exhibit deformation lamellae.

In many cases several grains have become joined to form mosaic-like aggregates. Grain to grain contacts in the mosaic are commonly indistinct, but where recognized have a sutured pattern. These aggregates are, in part, cemented by secondary silica (Plate 7, Fig. 3).

Lines of fluid inclusions occur in variable quantities throughout the quartz grains in these rocks. Many of these lines are restricted to single grains, but swarms of bubble trains may be continuous across several grains.

Besides quartzite fragments, chert, usually less than 1 percent, is the only rock fragment identified. Total rock fragment content seldom exceeds 2 to 3 percent.

The matrix consists of about equal amounts of sericite and crushed quartz. It occurs as scattered, irregular and isolated patches, or as thin films (mainly sericite) wedged between detrital grains. The average matrix content is about 3 percent; rarely a maximum of 5 or 6 percent is reached.

Authigenic silica, which occurs as overgrowths in optical continuity with detrital quartz grains, is the cement and may comprise up to 3 or 4 percent of the rock.

The heavy mineral content of these rocks is very low, averaging less than a percent, rarely attaining a level of 3 percent. The minerals present include: hematite, ilmenite, leucoxene, tourmaline, and zircon.

The upper Montalto, excluding the "blue quartzite" interval, is dominated by quartz. Quartz grains comprise about 90 percent of the rock on the average occurring as equant to subequant, subrounded grains. These grains range from about 0.2 - 1.5 mm in diameter; size sorting is fair.

Much of the quartz has combined to form granular mosaics. The aggregates, in which the quartz exhibits sutured or interlocked grain contacts, is cemented by secondary silica. Individual grains have intense undulose extinction, but deformation lamellae are not common.

TABLE 6. Point Count Data — Weverton Quartzite
(in percent)

Sample	Quartz	Microcline	Plagioclase	Micas	Rock Fragments	Crushed Quartz	Micromicas
1	34			0	13	30	33
2	49			0	4	14	33
3	46	1		0	4	8	41
4	63	1	2	1	1	12	20
5	38			1	32	9	20
6	69			0	13	9	9
7	54	1		0	7	12	26
8	89			0	1	11	
9	60			1	5	9	25
10	38	1		2	7	21	31
11	78		1	0	6	6	9
12	73			0	6	7	14
13	75			2	6	11	6
14	73	2	1	0	9	10	5
15	63	2	1	0	6	10	19
16	78			0	14	4	4
17	90	0	0	1	4	3	2

1 Southeast slope East Big Flat Ridge at elevation 1500′ about 0.5 mile south-east crest of Ram Hill.
2 Crest of Piney Mountain about 2.3 miles east of Caledonia Park.
3 Elevation 1460′ along Shippensburg Road.
4 Elevation 1180′ along Birch Run Road north of Wildcat Hill.
5 Float at elevation 1370′ along northwest slope Piney Mountain about 3.1 miles northeast of Caledonia Park.
6 Elevation 1610′ southeast crest of Ram Hill.
7 Southwest end of Grave Ridge.
8 South end of Piney Mountain about 3.0 miles east of Caledonia Park.
9 Elevation 1660′ about 0.8 mile north intersection Shippensburg Road and Route 233.
10 Southeast slope of Piney Mountain at elevation 1620′ about 0.3 mile south of Shippensburg Road.
11 Elevation 1630′, 3.4 miles east-northeast of Caledonia Park.
12 Elevation 1500′ along crest of Rocky Mountain.
13 Crest of Piney Mountain about 2.2 miles east of Caledonia Park.
14 Elevation 1180′ along Birch Run at south end of Big Pine Flat Ridge.
15 Elevation 1610′ along road north of Dead Woman Hollow.
16 Elevation 1790′ south of Dead Woman Hollow.
17 Near south end of East Big Flat Ridge at elevation 1730′.

Samples composed of abundant subequant quartz grains have a distinct dimensional orientation. Many of the oriented grains are partially bordered by thin zones of very finely comminuted quartz.

Only a few samples contain feldspar. The feldspar is mainly microcline but albite may also be present in very subordinate quantities. The grains are primarily anhedral laths which have undergone alteration in various degrees.

Rock fragments normally comprise less than 2 percent of the rock, and include quartzite and chert. Chert is less common than the quartzite fragments.

Interstices between grains are filled by a crushed quartz-sericite matrix and silica cement. The matrix generally makes up 1 to 4 percent of the rock; cement, about 0 to 3 percent. Together they do not exceed 5 percent of the rock.

Hematite, ilmenite, leucoxene, rutile, and zircon are the heavy minerals. These minerals constitute 1 to 2 percent of the rock; hematite is commonly the most abundant species.

Tourmaline, a very minor accessory, may have authigenic overgrowths with very irregular and angular terminations. This type of unabraded overgrowth is similar to that reported by Krynine (1946, p. 71-74) from other Cambrian sediments.

The "blue quartzite" zone is composed of rocks whose quartz content is between 60-85 percent. Grains of microcline (<1 percent), and quartzite and chert rock fragments (<1 percent) constitute the remainder of the grain content of the rock. The grains range in diameter from 0.1 - 2.0 mm, and the sorting is fair. The morphology and character of these components are similar to that exhibited in other parts of the Montalto Member of the Harpers Formation.

Moderately abundant matrix and authigenic silica plus considerable hematite are common in parts of the "blue quartzite" unit. Clay minerals, sericite, and rarely chlorite comprise the matrix which occurs in isolated, intergranular patches. The matrix forms about 7 to 10 percent of the rock.

Hematite is the second most abundant constituent in this unit. The hematite content is highly variable; it averages 3 to 7 percent but reaches 25 percent in some cases. This mineral may occur as individual grains or in irregularly-shaped masses. The non-granular variety which fills the interstices between grains or forms a thin rim completely or partly around quartz grains acts as a cement. In part, the hematite replaces the quartz. This relation is seen in both thin section and polished section (Plate 7, Fig. 4).

About 2 to 5 percent of the rock is authigenic silica, commonly occurring as overgrowths on detrital quartz. The original grain boundaries are usually outlined by very thin zones of hematite dust (Plate 7, Fig. 5).

The upper Harpers shows moderate size sorting. Grain size ranges from less than 0.06 mm to 1.0 mm in diameter, with an average of approximately 0.1 - 2.0 mm.

Mild to intensely strained, subangular, equant quartz and equant, anhedral feldspar are the most abundant grains. The quartz content is about 60-70 percent; feldspar, 3 to 7 percent (Plate 7, Fig. 6).

Both microcline and albite feldspar are present but microcline is far more abundant. Usually these grains show little alteration, although the margins may be partially corroded by reaction with the matrix, as are some quartz grains.

The remaining grains are primarily micaceous rock and chert fragments, and muscovite. These components do not exceed a few percent of the rock.

The matrix consists of crushed quartz, sericite and chlorite. The chlorite, notably absent from the matrix of other parts of the Harpers Formation, comprises half the matrix of the rocks in this unit. The matrix, occurring as isolated, irregular, intergranular patches, varies in amount from 15 - 25 percent.

Point count data for the Harpers is given in Table 7.

TABLE 7. Point Count Data — Harpers Formation
(in percent)

Sample	Quartz	Microcline	Plagioclase	Micas	Chlorite	Rock Fragments	Micromicas	Crushed Quartz	Cement
1	84	1				1	5	9	0
2	87					1	3	4	5
3	90			1		1	5	2	1
4	83					1	11	3	2
5	65	3	1	2	15	2	8	4	
6	48	3	2	2	9	1	27	8	
7	71	1				1	13	1	13
8	86					2	8	3	1
9	86					1	2	10	1
10	93					3	2		2
11	65		1			1	12	4	21
12	95					1	1	2	1

1 Montalto Member at elevation 1180′ on Birch Run Road.
2 Montalto Member northwest slope of Poke Hill.
3, 4 Float of Montalto Member from crest of Piney Mountain, about 3.2 miles west-southwest of Wenksville.
5 Upper Harpers base of Green Ridge at elevation 1480′.
6 Float of upper Harpers north of Rocky Knob.
7 Montalto "blue quartzite" east slope Green Ridge at elevation 1650′.
8 Montalto Member crest Piney Mountain about 3.5 miles west-southwest of Wenksville.
9 Float of Montalto Member crest of Yellow Ridge.
10 Montalto Member about 0.5 mile north Eagle Rock.
11 Montalto "blue quartzite" about 0.3 mile north of Wigwam Hill.
12 Montalto Member southeast slope Big Pine Flat at elevation 1770′.

Antietam Quartzite

Petrographic examination of the Antietam Quartzite shows that the rocks of this unit are basically well-sorted, medium-grained protoquartzites and quartzites.

The typical Antietam consists of 85-95 percent quartz occurring as detrital grains, quartzite rock fragments, crushed quartz, and silica cement. Quartz grains, which make up about 90 percent of the rock, are usually moderately to intensely strained. Deformation lamellae and peripheral granulation are also developed in many quartz grains (Plate 7, Figs. 6, 8).

Generally the quartz is equant and sub-rounded to rounded. However, in one sample in which cleavage is megascopically visible, these grains are subequant to elongate, and exhibit a recognizable dimensional orientation.

Rock fragments and feldspar are the only other constituents occurring as grains. Both are present in very subordinate amounts, generally not exceeding 3 percent.

Microcline is the identifiable feldspar although altered feldspar, which exhibits only albite twinning or shows alteration concentrated along cleavage traces without visible twinning, is also found. These grains may represent albite and orthoclase respectively.

Normally the feldspar is subhedral to anhedral and has the characteristic "grid-iron" twinning of microcline. The grains range from nearly fresh to moderately altered.

The chief rock fragments in the Antietam are quartzite, chert, and shale. Only quartzite fragments comprise as much as 1 percent of the rock.

TABLE 8. Point Count Data — Antietam Quartzite
(in percent)

Sample	Quartz	Microcline	Plagioclase	Rock Fragments	Micromicas	Crushed Quartz	Cement
1	77	2		1	12	4	4
2	75	2		1	11	5	1
3	88			2	2	5	3
4	81	1		1	1	1	15
5	89			1	1	1	8
6	89			2	2	4	3
7	81	2	1	1	9	4	2
8	88			1	1	4	6
9	83			3	1	2	11
10	90			1	1	3	5

1 Base of Wigwam Hill about 2.5 miles north-northeast of Caledonia Park.
2 North end (elevation 1300′) Quarry Gap Hill about 2.0 miles north of Caledonia Park.
3 North side Phillamen Run about 2.8 miles north of Fayetteville.
4 Near crest of Quarry Gap Hill about 1.8 miles north of Caledonia Park.
5 South end of Wolf Hill about 2.5 miles north-northeast of Caledonia Park.
6 Crest of Rocky Knob about 4.2 miles north-northeast of Caledonia Park.
7 North end Quarry Gap Hill (elevation 1660′) about 2.0 miles north of Caledonia Park.
8, 9, 10 Exposures in third level of Caledonia Sand Quarry about 2.2 miles west-south-west of Caledonia Park.

Crushed quartz, shredded muscovite, sericite, and illite comprise the matrix which fills the interstices between the grains. The amount of matrix in these rocks is highly variable and ranges from about 3 to 12 percent.

The micaceous material probably represents the recrystallization products of a previous argillaceous matrix, whereas the fine-grained quartz is principally the result of the granulation of the quartz grains.

The distribution of the matrix is generally spotty. Where the matrix is most abundant, the quartz is partially replaced by the matrix along the periphery of the grain, and crushed quartz is more common.

Silica, which occurs as authigenic overgrowths on detrital grains, is the main cementing material. Rarely, hematite occurs in local zones within a sample. The cement is 1 to 3 percent of the rock.

No carbonate has been recognized in these rocks although Nickelsen (1956, p. 251) reports up to 10 percent from northern Virginia. However the extremely friable and pitted weathered surface of some rocks would seem to suggest the presence of carbonate locally in the Caledonia area.

Heavy minerals may comprise 3 to 5 percent of the rock, although the average is less than 1 percent. Hematite and leucoxene are the most abundant and are usually accompanied by tourmaline, rutile, zircon, and ilmenite. Some of the tourmaline exhibits clear or pale colored authigenic overgrowths.

The mineral composition of a representative suite of Antietam samples is presented in Table 8.

APPENDIX B — GEOLOGY OF LOCAL AREAS

SOUTHERN PART OF PINEY MOUNTAIN

The structure of Piney Mountain is interpreted by Stose (1929, 1932) as the uniformly west-dipping limb of a normal anticline whose axis is located about two miles east of the study area on Bear Mountain and Big Hill.

Outcrops of Weverton Quartzite across the southern end of Piney Mountain exhibit a dip eastward except for local near-vertical dips. Cross-bedding and flow cleavage-bedding relations in the Weverton define alternate outcrop belts of normal and overturned strata. Piney Mountain is thus comprised of an overturned syncline on the east and an overturned anticline on the west (Plate 1, Cross Section C). These folds, on which smaller folds of similar geometry are superimposed, are part of the western limb of the Bear Mountain-Big Hill anticline.

Stose (1929, 1932) shows a normal sedimentary contact between the Weverton and the Montalto Formations at the western edge of Piney Mountain.

Strata of the upper middle Weverton, overturned and dipping at a moderate angle to the east, crop out near an exposure and float fragments of the lower Montalto. The intervening beds of upper Weverton are not present. The position of the westernmost Weverton to the “blue quartzite” unit of the Montalto is inconsistent with the normal stratigraphic separation between these beds. On this basis, a fault is postulated between the Weverton and Montalto. This structure appears to extend along the western edge of Piney Mountain from U. S. 30 northward for about five miles.

The association of this structure with the overturned limb of an anticline suggests that the fault is a moderate, eastward dipping thrust. Minimum stratigraphic offset is estimated at 500 feet.

ROCKY MOUNTAIN AREA

The structure cross sections of Stose (1909-1929) for the northern part of Rocky Mountain are revised in the light of graded bedding, cross-bedding, and flow cleavage-bedding relationships. This evidence consistently shows that the ridge-forming Weverton Quartzite is overturned steeply eastward on Rocky Mountain. Hence it is not the west limb of an upright anticline, but the overturned west limb of an anticline whose axis lies within the Catoctin volcanic belt to the east (Plate 1, Cross Section D).

The presence of the Loudoun Formation along the lower, eastern slope of the ridge is doubtful. Although this formation crops out on the north side of Route 233 at the southern end of Rocky Mountain, 4½ miles south of Caledonia Park, it is not known to occur within the map area.

Thus, the Weverton and Catoctin Formations are mapped in probable fault contact along the lower slopes of Rocky Mountain.

EAGLE ROCK - POKE HILL AREA

The area is part of the Grier Hollow anticline, a broad symmetric arch developed mainly on the Montalto Quartzite Member of the Harpers Formation. Moderate to steeply dipping Montalto and Antietam strata at the western rim of the structure form a double row of discontinuous ridges and knobs. These ridges are separated by a narrow, low valley carved in the less resistant rocks of the upper Harpers.

The culmination of the structure, located in the vicinity of Grier Hollow and Furnace Run, exposes Weverton Quartzite. Mapping reveals the anticline as a doubly-plunging fold whose southwestern extension divides into two low anticlines separated by a shallow syncline (Plate 1, Cross Sections B and C).

Faulting complicates the western portion of the fold. An east-northeast trending oblique fault causes a marked offset in the outer Antietam ridges at the southwest corner of the anticline. As fold axes and topography are shifted westward on the south side of the fault, the dislocation is interpreted as a right lateral strike-slip fault. Maximum horizontal displacement is approximately 2000 feet. A branch of the tear fault, located near the tributary to Phillamen Run, separates blocks of Antietam and Montalto outcrops which exhibit discordant structural orientation.

Along the western margin of the Grier Hollow anticline, the topographic separation of exposures of Montalto and Antietam is highly irregular, locally absent, and nowhere sufficient to accommodate a full thickness of upper Harpers. On this evidence, a steep longitudinal reverse fault is mapped extending from the major tear fault (near Eagle Rock) northeast to the vicinity of Poke Hill.

Although Stose (1909) recognized the Grier Hollow anticline, neither the core of Weverton Quartzite nor the flank faults mapped by the present writer were identified by him.

LONG PINE RUN - BIRCH RUN AREA

Just east of Big Flat Ridge the Antietam Quartzite outlines a highly compressed, southwest plunging syncline. The Tomstown Dolomite is inferred along the axis of the structure on the basis of the adjacent stratigraphy, iron-ore pits (a common phenomenon in basal Tomstown in this region), and shallow, marshy depressions which are visible in the field and on air photographs and are interpreted as solution features.

The strata of the syncline dip east; the western limb dips at a moderate angle and the eastern limb is steeply overturned. The overturned limb is incomplete owing to faulting.

Stose (1929, 1932, 1953) terms the bordering fault the Cold Spring thrust, retaining the name of a similar fault mapped in the Carlisle quadrangle to the northeast. Although actual physical continuity between the two structures is not confirmed by published field investigations, the writer uses the name Cold Spring thrust.

The map pattern and character of the Hosack Mine syncline follows closely the interpretation of Stose (1929, 1932) as far north as Long Pine Run. North of this locality, the syncline is disrupted by cross-faults not previously mapped. The location of the "blue quartzite" unit was omitted from the outcrop of the upper Montalto on the south limb of this syncline (Plate 1). Figure 18 shows its correct position.

Two fault-bounded slices occur in Birch Run Valley north of Long Pine Run. These zones, one enclosing Montalto and the other Antietam, are bordered on the east by the Hosack thrust and to the west by secondary thrusts.

North of Rocky Knob the formations involved in the Hosack Mine syncline "wrap around" the end of the fold. The offset of mapped units and the deflection of bedding-cleavage intersections (L_{ox1}) in this area indicate that the axis of the structure is offset along two prominent cross-faults. The Rocky Knob fault and a subsidiary fault to the south of the valley of Knob Run have displaced the northernmost end of the fold to the east as much as 2000 feet. In view of this lateral movement, the axial trace of the Hosack Mine syncline must lie near the crest of East Big Flat Ridge rather than in the valley of Birch Run as previously mapped (Stose, 1929).

As a result of the faulting, the quartzites which crop out along the western crest of East Big Flat Ridge in the vicinity of Strasbaugh Hill must be Antietam rather than Montalto (Stose, 1929). This interpretation is also supported by the location and distribution of the Montalto "blue quartzite" unit.

The Rocky Knob fault offsets the Cold Spring thrust east of Rocky Knob. The trace of the thrust is located about 1500 feet east of its previously mapped position south of Strasbaugh Hill (Stose, 1929). In this area, the upper part of the Montalto has been thrust northwestward, concealing the upper Harpers and parts of the basal Antietam in the overturned, eastern limb of the syncline.

Northwestward, the Cold Spring fault cuts across the axis of the syncline, and west of Ram Hill displaces slightly the "blue quartzite" on the normal limb of the fold.

CONOCOCHEAGUE CREEK VALLEY

The valley occupied by Conococheague Creek is a southwest plunging syncline faulted on the west. The outline of the fold is defined by the mapped trace of the "blue quartzite" unit of the Montalto.

Traced northeastward in float fragments, the "blue quartzite" occurs along the lower western slope of Piney Mountain for a distance of about five miles north of

Caledonia Park. This eastern limb of the fold, as shown by offset in the pattern of the "blue quartzite," is transected by small faults. The faults, located near the northern end of the Chambersburg Reservoir and about 4000 feet further north, define a segment of "blue quartzite" which has moved relatively westward. Horizontal displacement is about 200-500 feet.

The "blue quartzite" at the northern end of the syncline, is located about 1000 feet south of the junction of Route 233 and the Miltonberger Road, and the Rocky Knob cross-fault.

Fragments of the "blue quartzite" unit on the western flank of the structure trend oblique to the adjacent formations on East Big Flat Ridge and the longitudinal fault followed by Conococheague Creek. About 4000 feet from the nose of the syncline, the west limb is truncated by the fault.

For much of its length south of the Rocky Knob cross-fault, the Laurel Forge fault juxtaposes upper Montalto on the east against lower Montalto. Locally, as at Hosack Run, Weverton Quartzite occurs west of the fault. Nowhere along the Laurel Forge fault nor the axis of the adjacent Wildcat Hill anticline are strata of the Loudoun Formation exposed. It is probable that Stose (1929) misidentified beds of the lower middle Weverton as Loudoun in this area.

The Caledonia Park syncline is on the down-dropped side of the fault. Because the Laurel Forge fault is not exposed, its dip cannot be observed directly, and it is not known whether it is reverse or normal in character. In view of its straight trace, the fault must have a high angle of dip. A locality where the "blue quartzite" is juxtaposed with the base of the lower Montalto provides the best estimate of stratigraphic displacement. Offset is about 1200-1500 feet.

PINEY MOUNTAIN WEST OF STRASBAUGH SCHOOL

The topographic re-entrant on Piney Mountain one mile west-northwest of Strasbaugh School occurs on the axial portion of the Piney Mountain anticline and exposes the phyllite member of the Loudoun Formation. The short knob east of the re-entrant and the continuation of Piney Mountain on the west, underlain by Weverton Quartzite, are on the limbs of the structure. Stose (1929) mapped the presence of Loudoun and the anticlinal character of this area, but other structures are not recognized.

The east flank of the anticline is a tight synclinal fold in the Weverton Quartzite truncated on the west by a small longitudinal fault (Plate 1, Cross Section B). Steeply dipping Loudoun conglomerate is exposed on the east limb of the syncline but is absent on the nose of the fold and the western limb probably as a result of faulting.

Loudoun phyllites exposed at the center of the structure dip beneath the younger and topographically higher Weverton Quartzite. Dips of 10-15 degrees southwest are common in the Weverton and approximate the plunge of the fold.

West of the anticlinal axis on the upper slopes of Piney Mountain, gray phyllites of the Loudoun Formation are unusually well exposed. These strata display excellent development of flow cleavage (S_1), slip cleavage (S_2'), and associated chevron folds. The intersection of the two cleavages ($L_{1\times2}'$) and the axes of the folds have an orientation which conforms to the axis of the bedding fold.

Loudoun phyllite also crops out near the crest of Piney Mountain on the western limb of the structure. Abundant phyllite fragments are found west of the crest and may be related to smaller order folds developed on the steep west limb of the Piney Mountain anticline. Because the Loudoun exposures on the ridge crest strike north

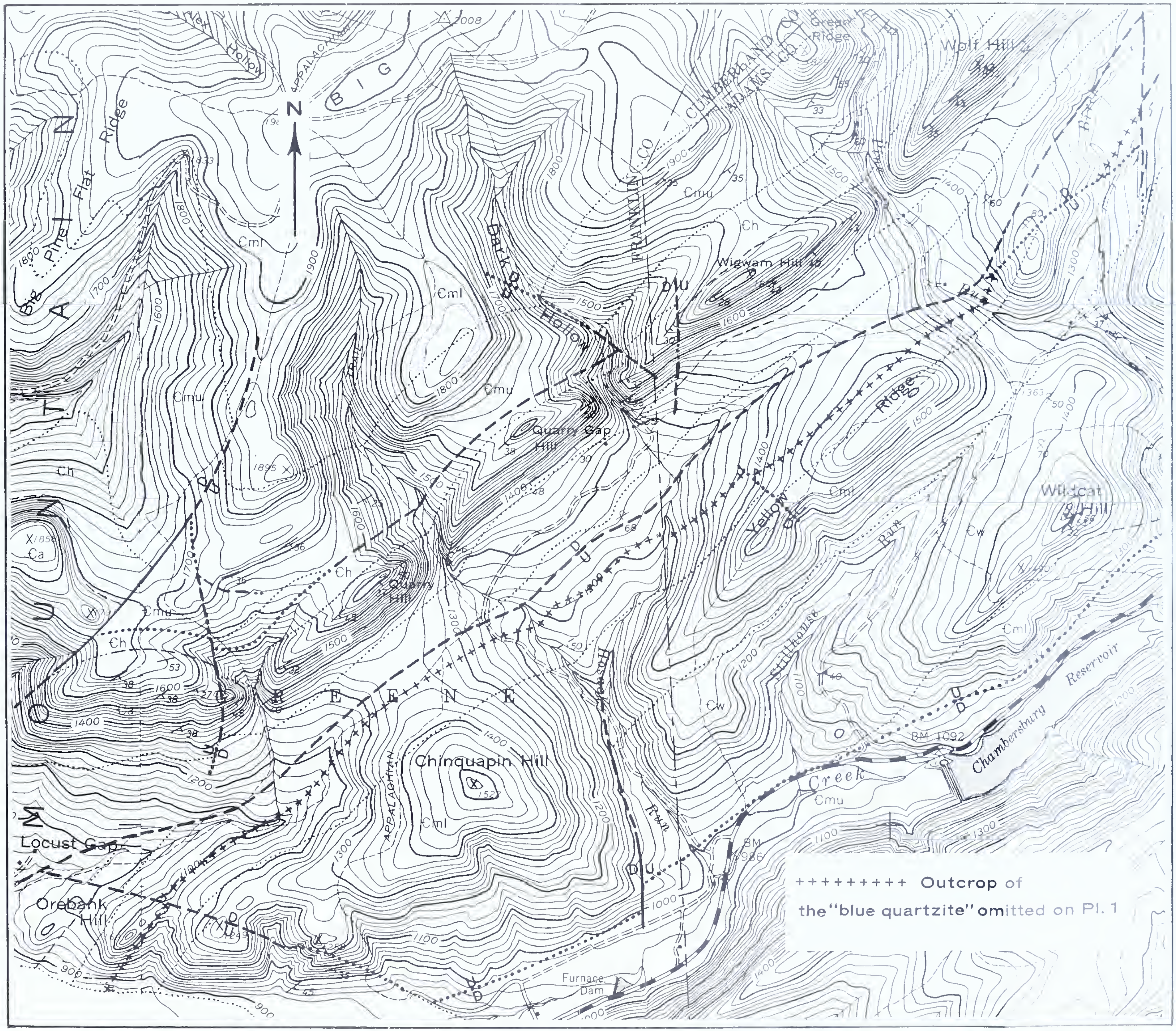

Figure 18. Geology of the Chinquapin Hill-Yellow Ridge area, showing the outcrop of the "blue quartzite."

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and directly into the Weverton Quartzite, a fault is mapped. The fault trends west-northwest across the summit at an elevation of 1720 feet. The west end of this oblique fault may merge with the Piney Mountain thrust. An alternate interpretation would show the diagonal fault offsetting the thrust, and then connecting with the northernmost fault on the east limb of the Caledonia Park syncline which shows the same strike and similar sense of movement.

The strike of the fault changes to northeast on the eastern slope of Piney Mountain, where the fault accounts for the contact of Weverton Quartzite and Catoctin volcanics. It is intersected by the Rocky Knob fault about 1.6 miles north of Strasbaugh School. North of the cross-fault, the thrust cannot be identified between the Loudoun Formation and Catoctin volcanics.

The Weverton and Montalto may be in fault contact on the western slope of Piney Mountain north of the Loudoun exposures. This fault, possibly a branch of the Piney Mountain thrust, is mapped on the basis of the closeness of the Weverton and the "blue quartzite" unit, the apparent absence of the upper Weverton, and the lack of a dip reversal in the Weverton west of the ridge crest.

MOUNTAIN CREEK VALLEY

Stose (1929, 1932) describes the structure in the area of Mountain Creek as a syncline enclosing the Antietam and Tomstown Formations whose western limb is cut off by the Laurel Forge fault. The Mountain Creek syncline exists as a mappable structure northeast along its plunge into the Carlisle quadrangle.

The Laurel Forge fault and the western margin of the syncline are mapped as parallel structures by Stose (1929, 1932). However, topographic features and the strike of bedding is oblique to the fault. These data necessitate a revision of the stratigraphy and structural pattern in the Mountain Creek valley.

The orientation of bedding on the knob at the east side of the syncline and near the northern border of the area is recognizable by the orientation of *Skolithos* tubes and silty laminae. The strike of the beds outlines an abrupt swing of the Antietam to the west-northwest and toward Grave Ridge. The Antietam on the west side of the syncline strikes southeast and projects into the east limb about one mile south of the Adams County line on Route 233.

The location of the nose of the Mountain Creek syncline in the Antietam at this point sharply contrasts with Stose's map (1929) which shows the Antietam extending about 1¼ miles farther south.

The writer infers that the low area in the Mountain Creek valley is not carved in the Tomstown Dolomite (Stose, 1929, 1932) but has developed on the relatively weak rocks of the upper Harpers. This interpretation is supported by exposures of gentle, east-dipping Montalto near the southern end of Grave Ridge striking obliquely into the Laurel Forge thrust, and the occurrence of characteristic cross-bedded lower Montalto farther southwest. This pattern of truncation of successively older rocks by the Laurel Forge fault on the west side of the Mountain Creek syncline is consistent with structural and stratigraphic evidence as well as the linear topography.

On the east limb of the syncline, the proximity and strike of the Antietam and Montalto require a fault. This fault and other smaller dislocations toward the center of the syncline may account for the large areal extent of the upper Harpers.

The Laurel Forge thrust along the western margin of the syncline divides to enclose wedges of Weverton Quartzite locally. The general trace of the fault, however, is nearly straight and indicates a high angle of dip.

Stose has at various times described the Laurel Forge fault as a normal fault (1929, 1932) and thrust (1953). Most recently, on the basis of mapping in the Carlisle quadrangle to the northeast, he described the fault as a diagonal structure which has juxtaposed Catoctin volcanics and Cambrian quartzose rocks on the northwest and Tomstown Dolomite on the southeast: ". . . this is probably due to thrusting of the syncline (Mountain Creek syncline) of Tomstown Dolomite northwest over the Precambrian rocks in the anticline . . ." (Stose, 1953).

For almost its entire length the Laurel Forge fault places younger rocks on the east in contact with older rocks on the west. If the fault is assumed to dip steeply east as indicated by Stose (1953), the map pattern of the structure and the attitude of the beds cannot be easily reconciled with thrusting. A normal fault is a simpler explanation.

Along Grave Ridge, the Antietam is in fault contact with the Weverton Quartzite which is located on the east limb of the Hammond Rocks anticline. Assuming a steep east or west dip, the estimated dip slip is about 4000 feet for a fault which shows about 2000 feet of stratigraphic separation.

PLATES

PLATE 2. — LOWER CAMBRIAN SANDSTONES AND CONGLOMERATES: BEDDING AND CLEAVAGE IN CAMBRIAN AND ORDOVICIAN ROCKS

Figure 1. — Loudoun Formation (conglomerate member). Interbedded polymictic conglomerate and quartzose phyllite southeast slope Piney Mountain, north Shippensburg Road.

Figure 2. — Loudoun Formation (conglomerate member). Rounded quartz and rhyolite pebbles (A) and flattened phyllite fragments (B) in graywacke matrix. Six inch scale. Locality as in Figure 1.

Figure 3. — Weverton Quartzite. Cross-bedded, bluish and grayish phyllitic graywacke of basal Weverton; Piney Mountain, northwest of Strasbaugh School.

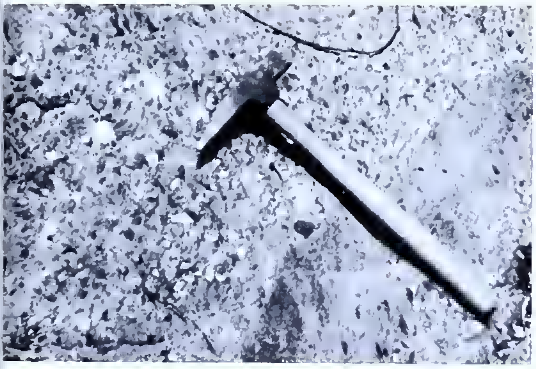
Figure 4. — Weverton Quartzite. Bluish and greenish gray, banded protoquartzite of upper middle unit. Bedding dips west, flow cleavage east. Piney Mountain southwest of Wenksville.

Figure 5. — Antietam Quartzite. Vertical, *Skolithos*-bearing quartzite Devil Alex Hollow. *Skolithos* normal to bedding.

Figure 6. — Conococheague Formation. Gray limestone, laminae and bands of dolomitic limestone in positive relief. Hammer on flow cleavage plane. Locality northeast Scotland.

Figure 7. — Weverton Quartzite. Bedding vertical, flow cleavage dips southeast. Locality south end Piney Mountain.

Figure 8. — Conococheague Formation. East-dipping flow cleavage, vertical bedding (A). Details of cleavage-lithology relations (B). Northeast of Scotland.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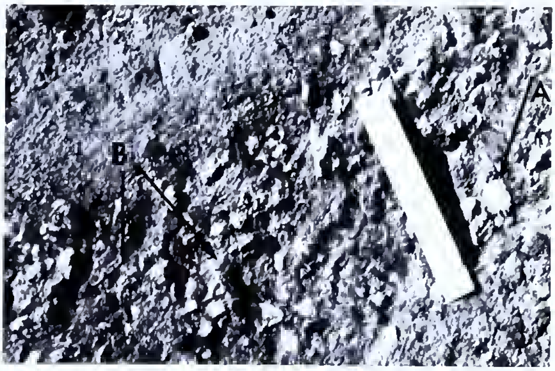

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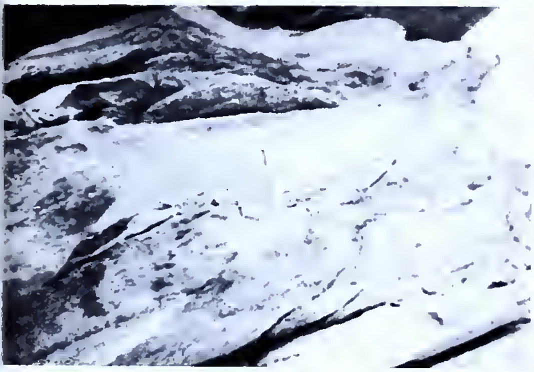
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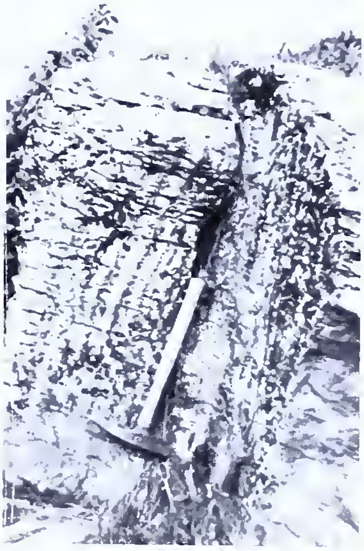
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PLATE 3. — PHOTOMICROGRAPHS OF GRAINS ELONGATED PARALLEL TO FLOW CLEAVAGE; CLEAVAGE

Figure 1. — Catoctin Formation. (Red porphyritic metarhyolite) elongate micropoikilitic groundmass parallels flow cleavage (x35, x-nicols).

Figure 2. — Elbrook Formation. Oölitic limestone. Oölites elongated parallel to flow cleavage (x35, x-nicols).

Figure 3. — Weverton Quartzite. Quartz grains with "tails" of crushed quartz oriented parallel to aligned matrix and flow cleavage (x50, x-nicols).

Figure 4. — Weverton Quartzite. Peripheral granulation and crushed quartz. Crushed quartz "tails", aligned matrix, and flow cleavage are parallel (x70, x-nicols).

Figure 5. — Folded flow cleavage in Loudoun phyllite. Slip cleavage (A) is axial plane of chevron fold. Scale is six inches.

Figure 6. — Large chevron fold developed in quartzose phyllite of Weverton. Slip cleavage formed in axial region of fold (A); quartz vein (B) parallels folded flow cleavage.

Figure 7. — Sub-horizontal fracture cleavage in folded argillaceous shales of Elbrook Formation near Conococheague Quarry.

Figure 8. — Flow cleavage in Catoctin Formation (Mottled metarhyolite) dips southeast. Dark streaks are aligned downdip in cleavage plane. Six inch scale.

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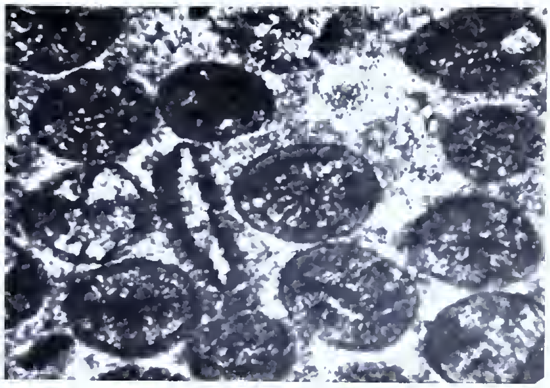
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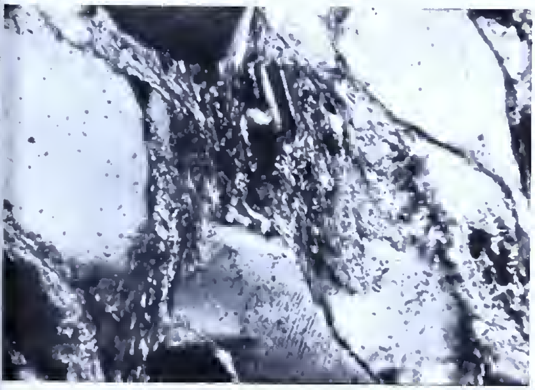
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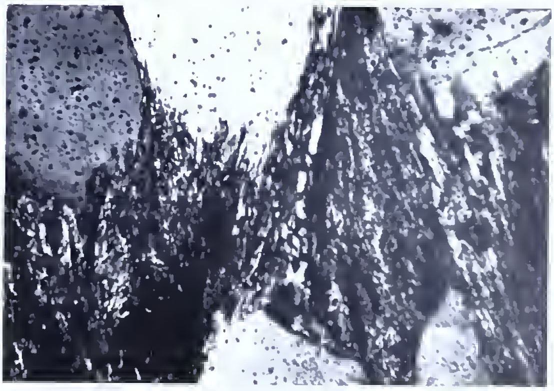
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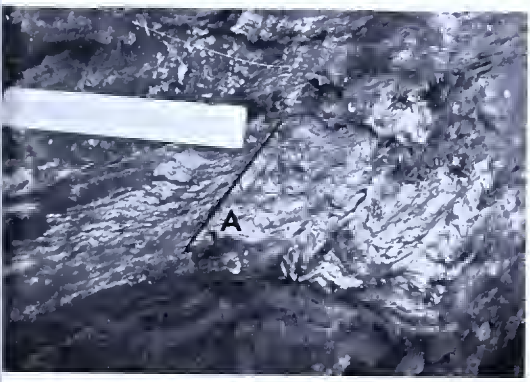

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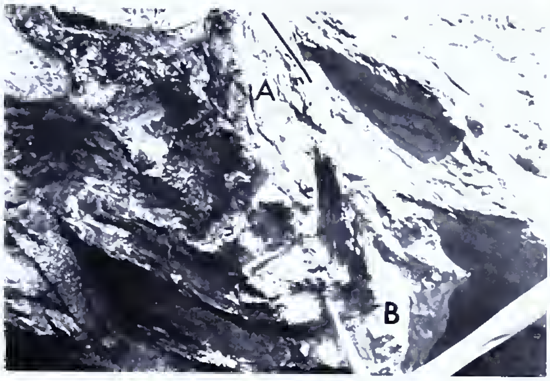

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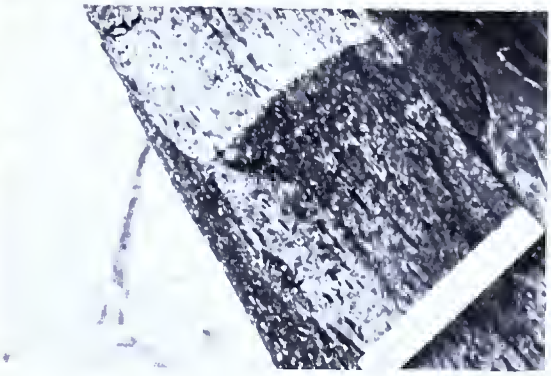
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PLATE 4. —BEDDING FOLDS: QUARTZ GRAIN DEFORMATION IN ANTIETAM QUARTZITE

Figure 1. — Trough of overturned syncline developed in Weverton Quartzite, north side Dead Woman Hollow.

Figure 2. — Offset of flow cleavage in phyllitic bed caused by flexural slip of adjacent, more competent sandstones in overturned limb of syncline. Locality as in Figure 1.

Figure 3. — Asymmetrical syncline in Antietam Quartzite, Mt. Cydonia Quarry.

Figure 4. — Crestal portion of anticline in Antietam Quartzite, Caledonia Quarry.

Figure 5. — Recrystallized quartz (x50, x-nicols).

Figure 6. — Strained quartz grains; strain shadows (A) and deformation lamellae (B) (x60, x-nicols).

Figure 7. — Detrital quartz with overgrowth in optical continuity with grain (x60, x-nicols).

Figure 8. — Intensely strained detrital quartz; strained shadows (A) and deformation lamellae (B) shown characteristic perpendicular relationship (x100, x-nicols).

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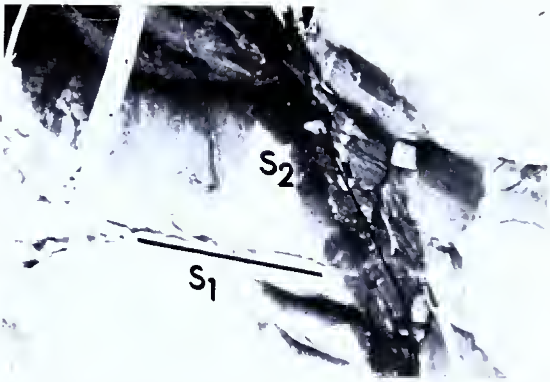

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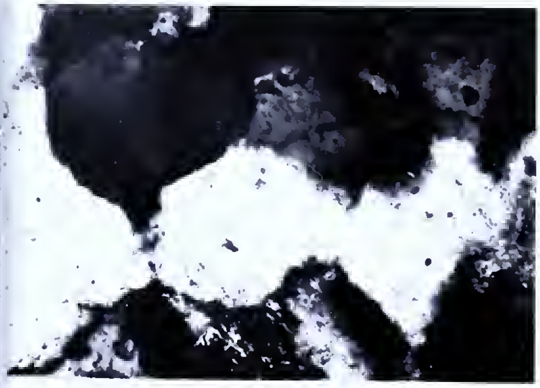

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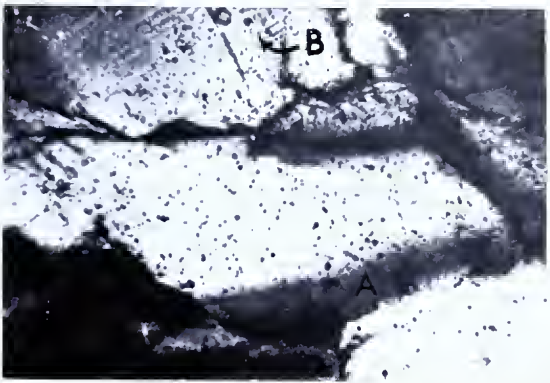

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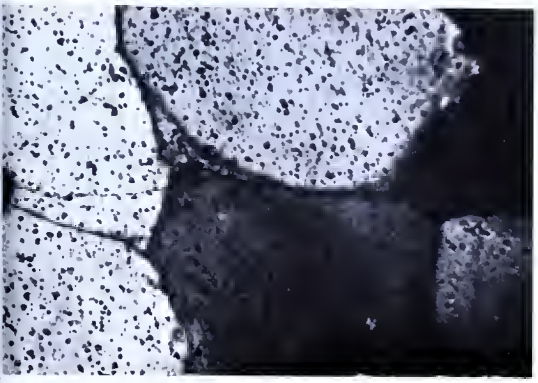

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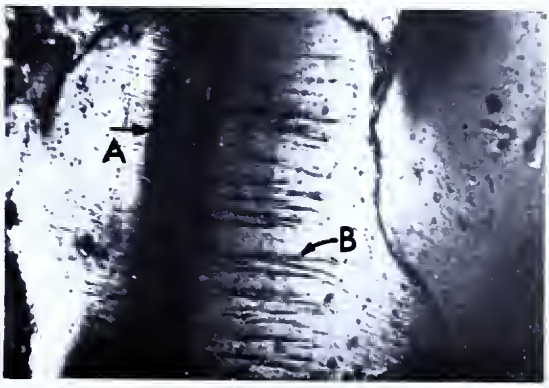

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PLATE 5. — MINOR BEDDING AND CLEAVAGE FOLDS AND PHOTOMICROGRAPHS OF THE CATOCTIN METABASALT

Figure 1. — Axial portion minor recumbent fold in Elbrook Formation near Conococheague Quarry. Thin bedded, shaly limestone has horizontal fracture cleavage (at hammer head).

Figure 2. — Small, recumbent shear folds with axial plane cleavage in Elbrook Formation east of Scotland.

Figure 3. — Intersecting flow cleavage (A) and slip cleavage (B) in Loudoun phyllite (x35).

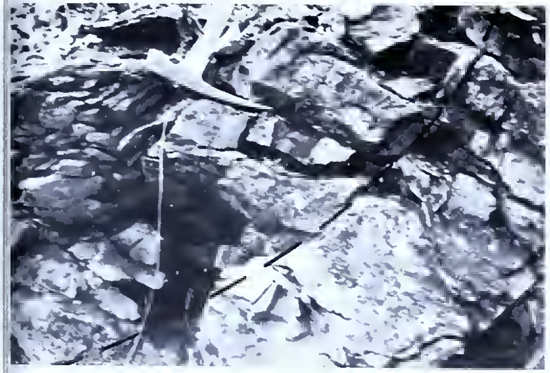
Figure 4. — Shear on slip cleavage has dragged micas into slip cleavage plane (A) the axial region of folded flow cleavage. Flow cleavage outlined by hematite (black), quartz grains (B) and aligned micas (x100, x-nicols).

Figure 5. — Fine grained metabasalt. Albite microlites outline felty texture (x35, x-nicols).

Figure 6. — Porphyritic metabasalt. Combined pericline, albite, and Carlsbad twinning in albite phenocryst (x35, x-nicols).

Figure 7. — Porphyritic metabasalt. Albite phenocryst with irregular zones of low birefringent, cryptocrystalline material (A); minor alteration along margin of crystal (x35, x-nicols).

Figure 8. — Amygdaloidal metabasalt. Albite crystals and core of vermicular chlorite fill amygdule (x35, x-nicols).



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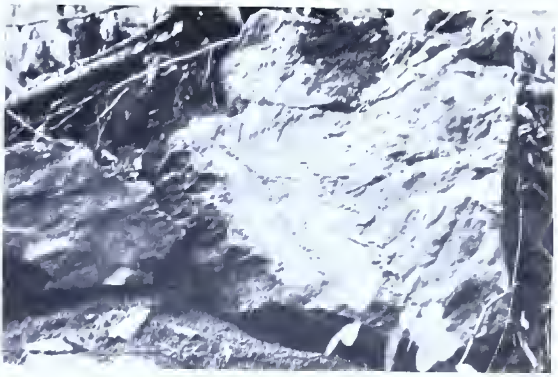

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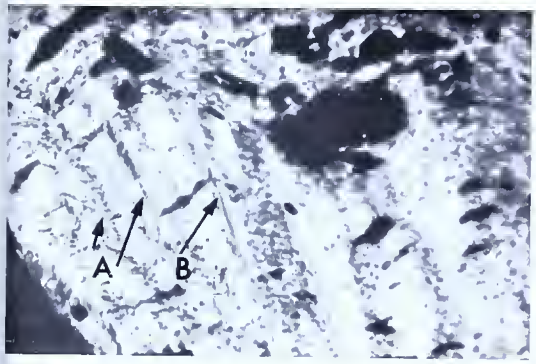

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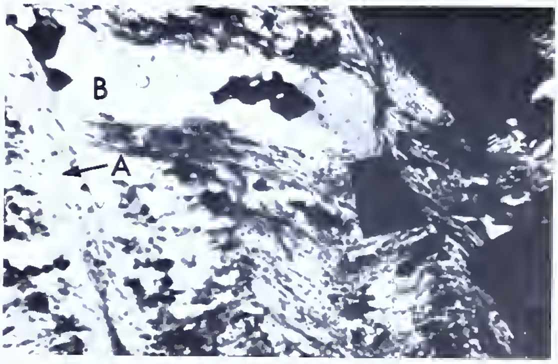

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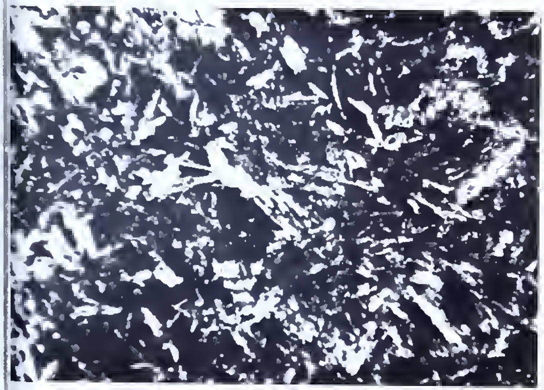

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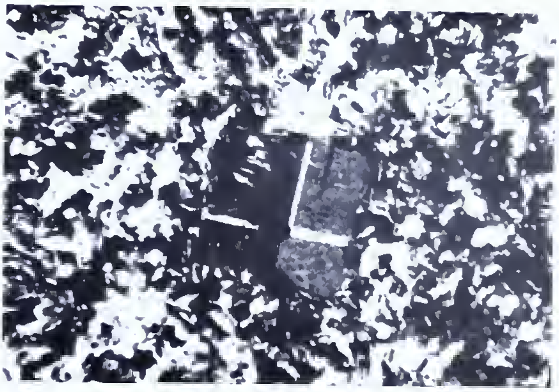

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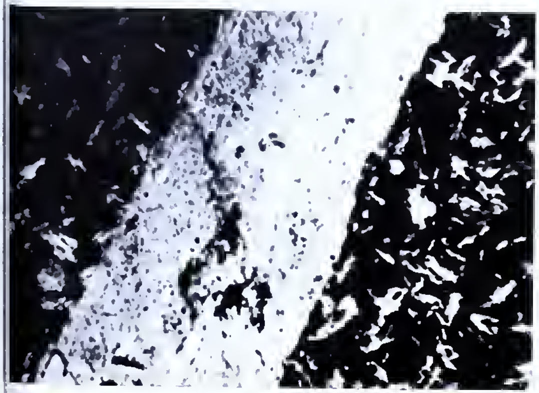

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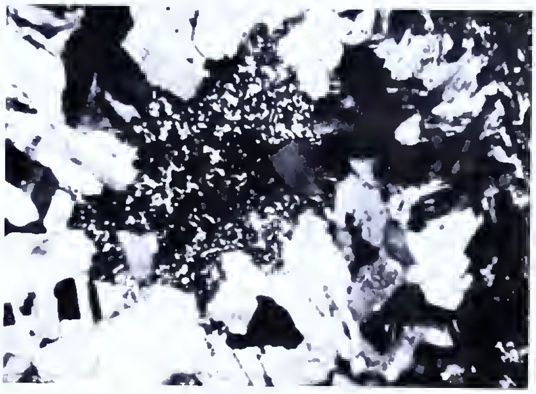

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PLATE 6. — PHOTOMICROGRAPHS OF THE CATOCTIN METARHYOLITE AND LOUDOUN FORMATION

Figure 1. — Red porphyritic metarhyolite. Albite phenocryst with characteristic chessboard twin structure in micropoikilitic groundmass.

Figure 2. — Blue metarhyolite. Strained feldspar phenocryst segmented by tension fractures oriented normal to flow cleavage in hand sample. Fracture filling is fine grained quartz and sericite (x35).

Figure 3. — Mottled metarhyolite. Quartz phenocryst set in undeformed micropoikilitic groundmass (x35, x-nicols).

Figure 4. — Metarhyolite breccia. Prismatic crystals of piedmontite between rhyolite fragments (x35).

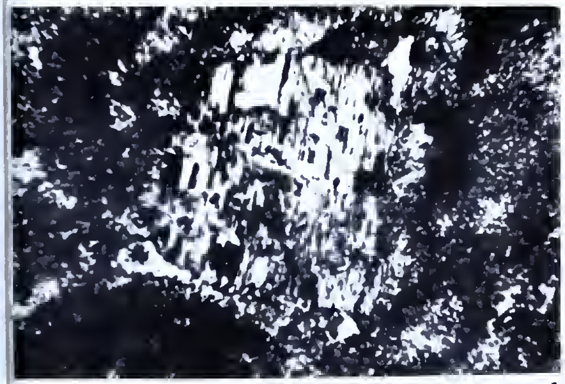
Figure 5. — Mottled metarhyolite. Highly strained quartz phenocryst and sheared micropoikilitic groundmass (x35, x-nicols).

Figure 6. — Mottled metarhyolite. Aligned micropoikilitic groundmass warped by rotation of feldspar phenocryst (x35, x-nicols).

Figure 7. — Loudoun Formation. Muscovite-sericite-quartz phyllite (x35, x-nicols).

Figure 8. — Loudoun conglomerate. Quartz and phyllitic rock fragments in an aligned quartz-sericite groundmass (x35, x-nicols).

PLATE 6



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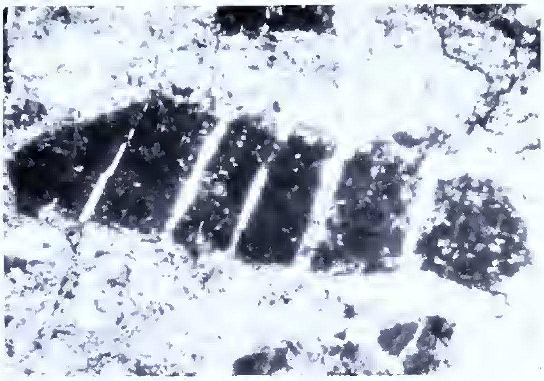

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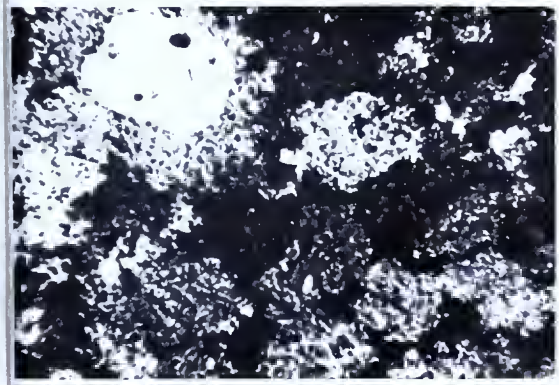

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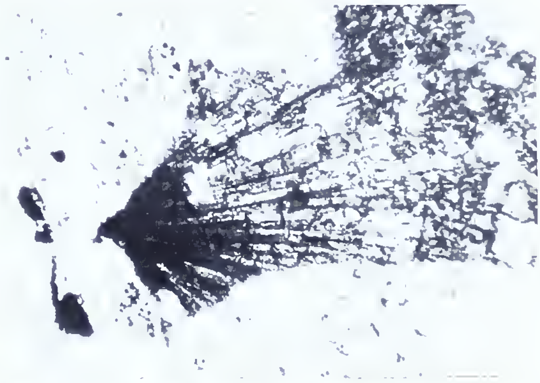

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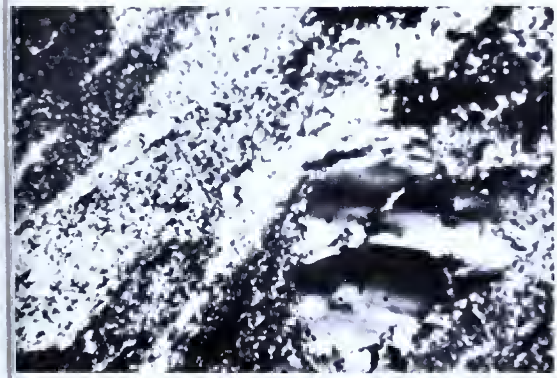

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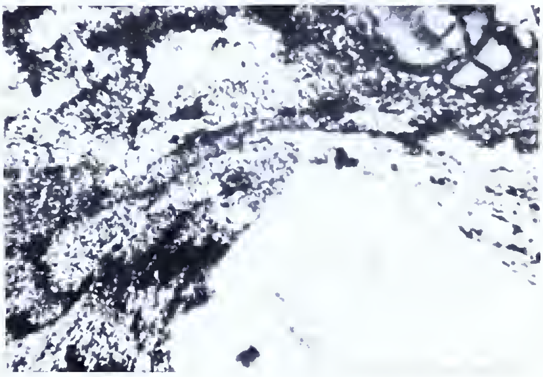

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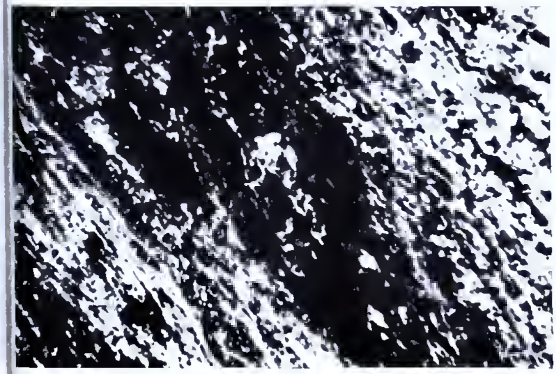

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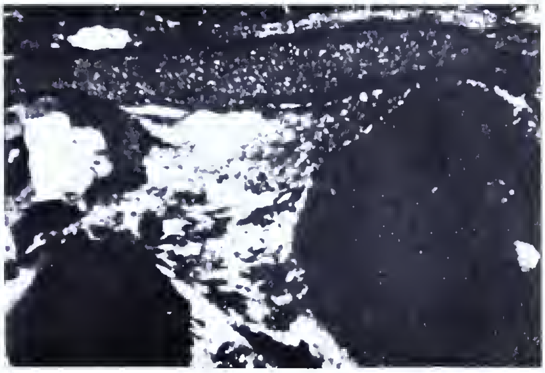

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PLATE 7. — PHOTOMICROGRAPHS OF THE WEVERTON QUARTZITE, MONTALTO QUARTZITE, HARPERS FORMATION AND ANTIETAM QUARTZITE

Figure 1. — Weverton Quartzite (lower unit). Strained quartz and microcline in a quartz-sericite matrix. Quartz grains are dimensionally aligned (x35, x-nicols).

Figure 2. — Weverton Quartzite (lower middle unit). Subangular quartz grains exhibit undulose extinction (A) and deformation lamellae (B). Quartz oriented parallel to aligned matrix (x35, x-nicols).

Figure 3. — Montalto Quartzite (lower unit). Mosaic of recrystallized detrital quartz. Matrix is finely granulated quartz (x35).

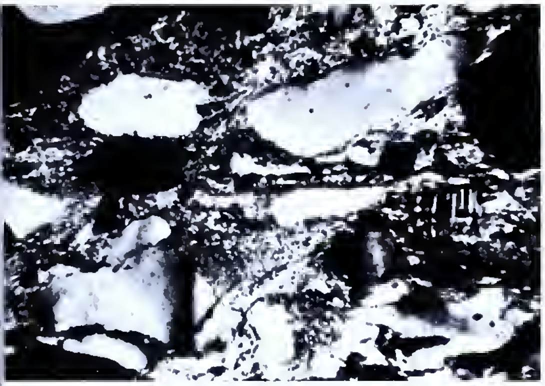
Figure 4. — Montalto "blue quartzite". Hematite-cemented quartzite consisting of subangular to subrounded, equant detrital quartz (x35).

Figure 5. — Montalto Quartzite (upper unit). Authigenic overgrowth in optical continuity with detrital quartz grains. Border of original grain marked by hematite (x35, x-nicols).

Figure 6. — Harpers Formation (upper Harpers). Low-rank graywacke composed of quartz, plagioclase, and microcline in a matrix of chlorite, crushed quartz, and sericite. (x35, x-nicols).

Figure 7. — Antietam Quartzite. Subrounded, equant quartz grains with strain shadows and deformation lamellae. Matrix is mainly sericite (x35, x-nicols).

Figure 8. — Antietam Quartzite. Strain shadows (A) and deformation lamellae (B) in partially recrystallized orthoquartzite (x35, x-nicols).



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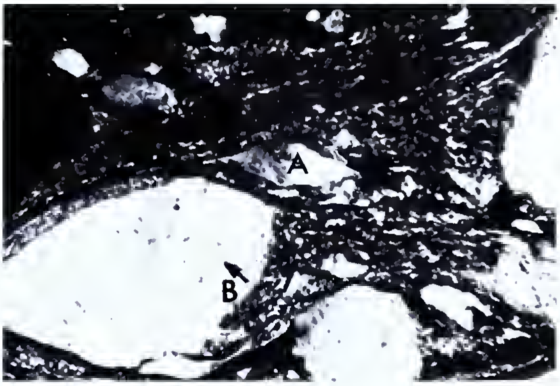

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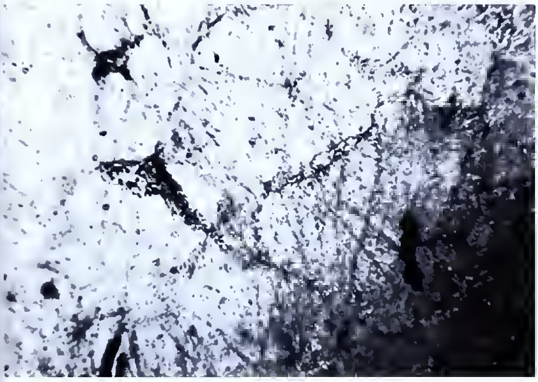

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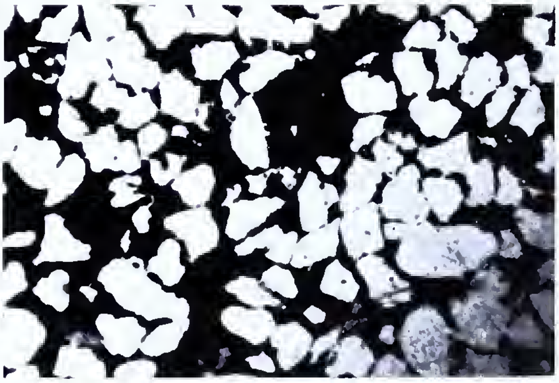

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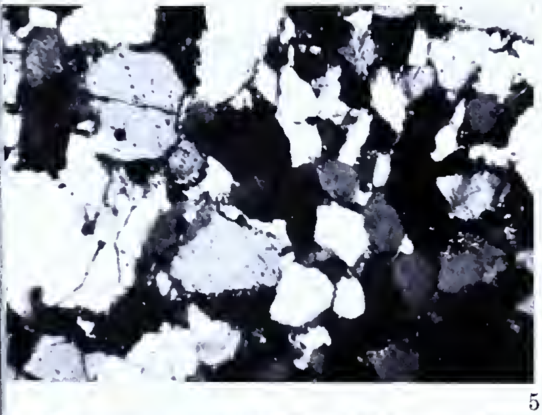

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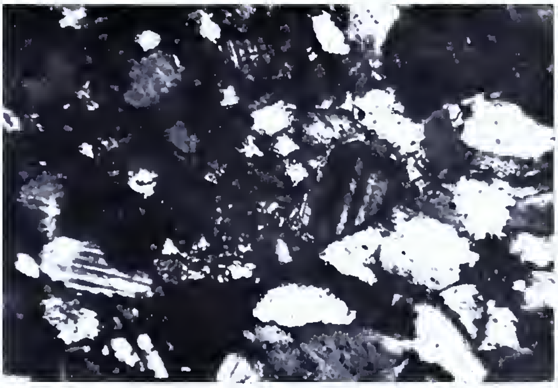

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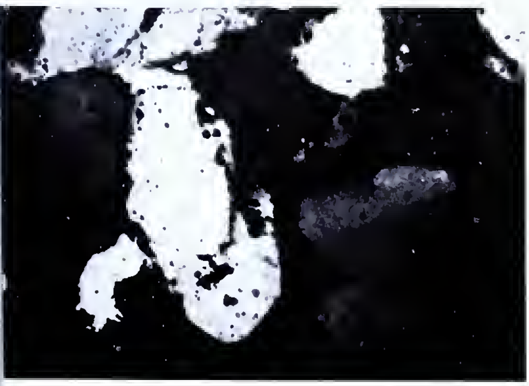

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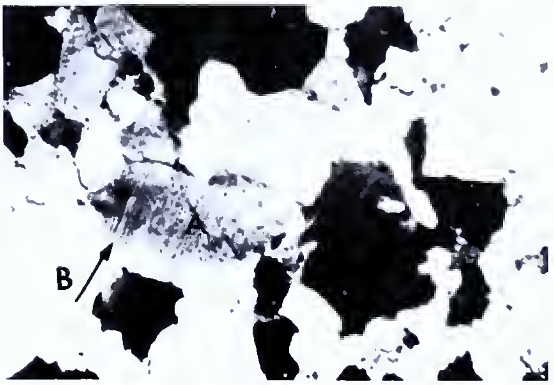

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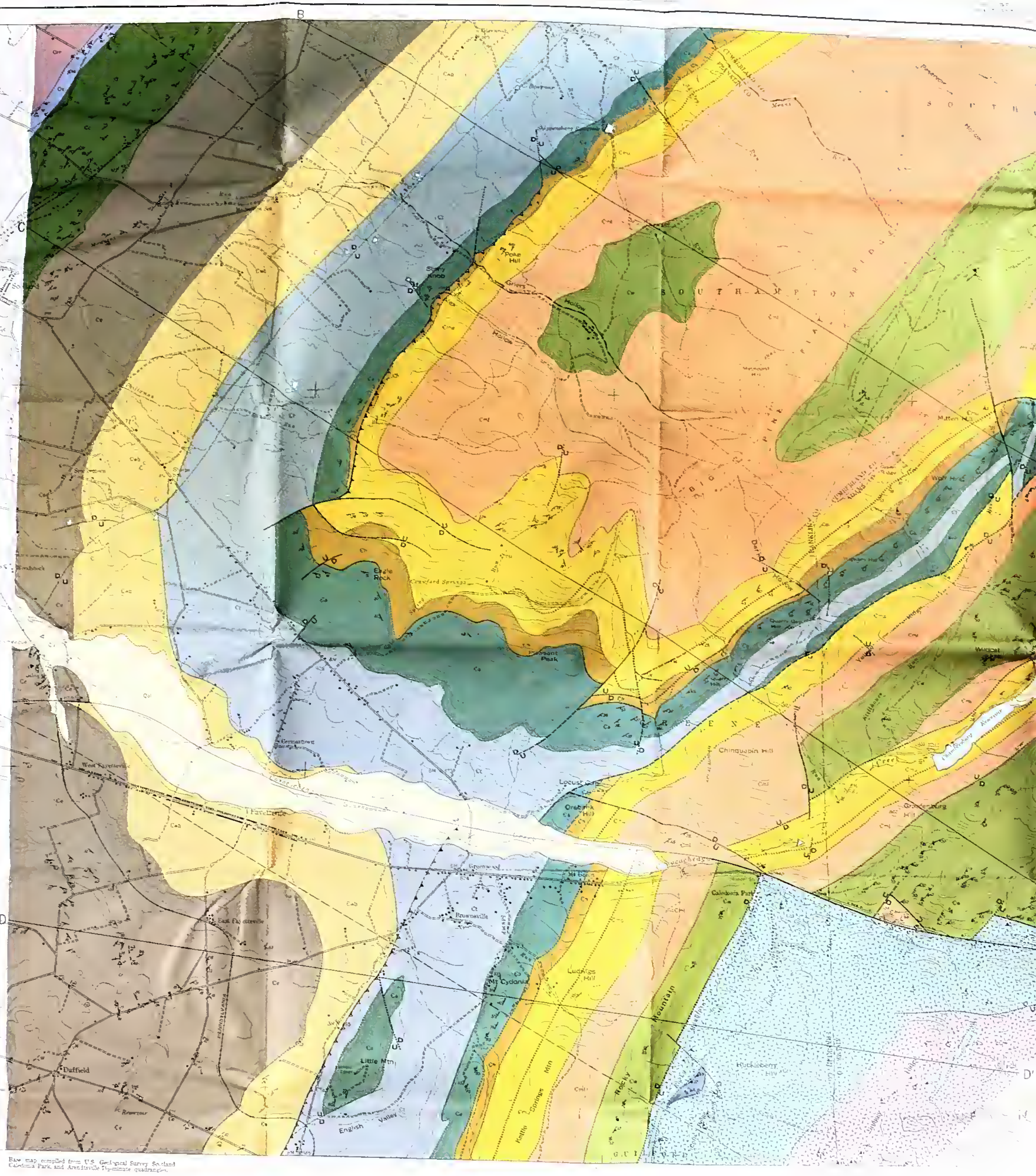
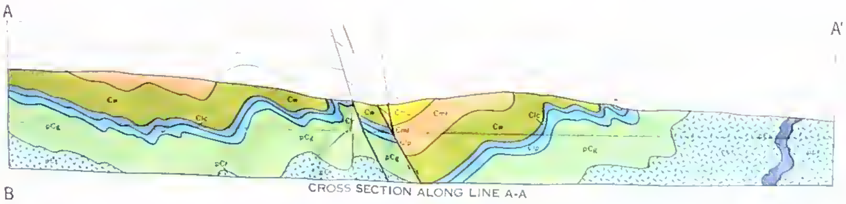


Base map compiled from U.S. Geological Survey Scotland, Caledonia Park, and Arendtsville 7½-minute quadrangles.

CROSS SECTIONS



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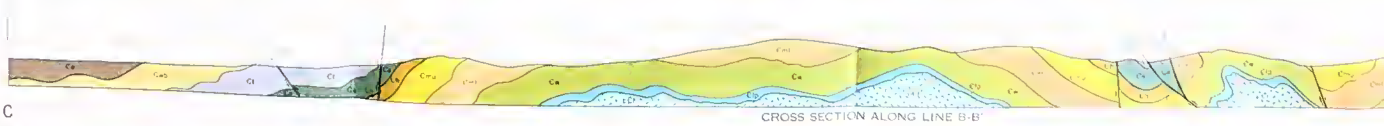

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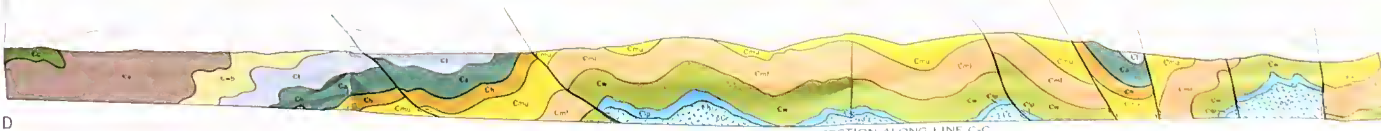

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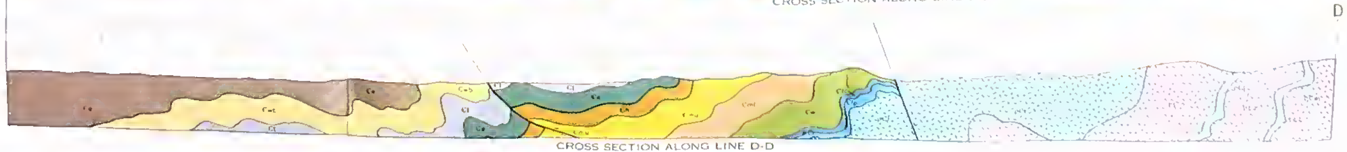

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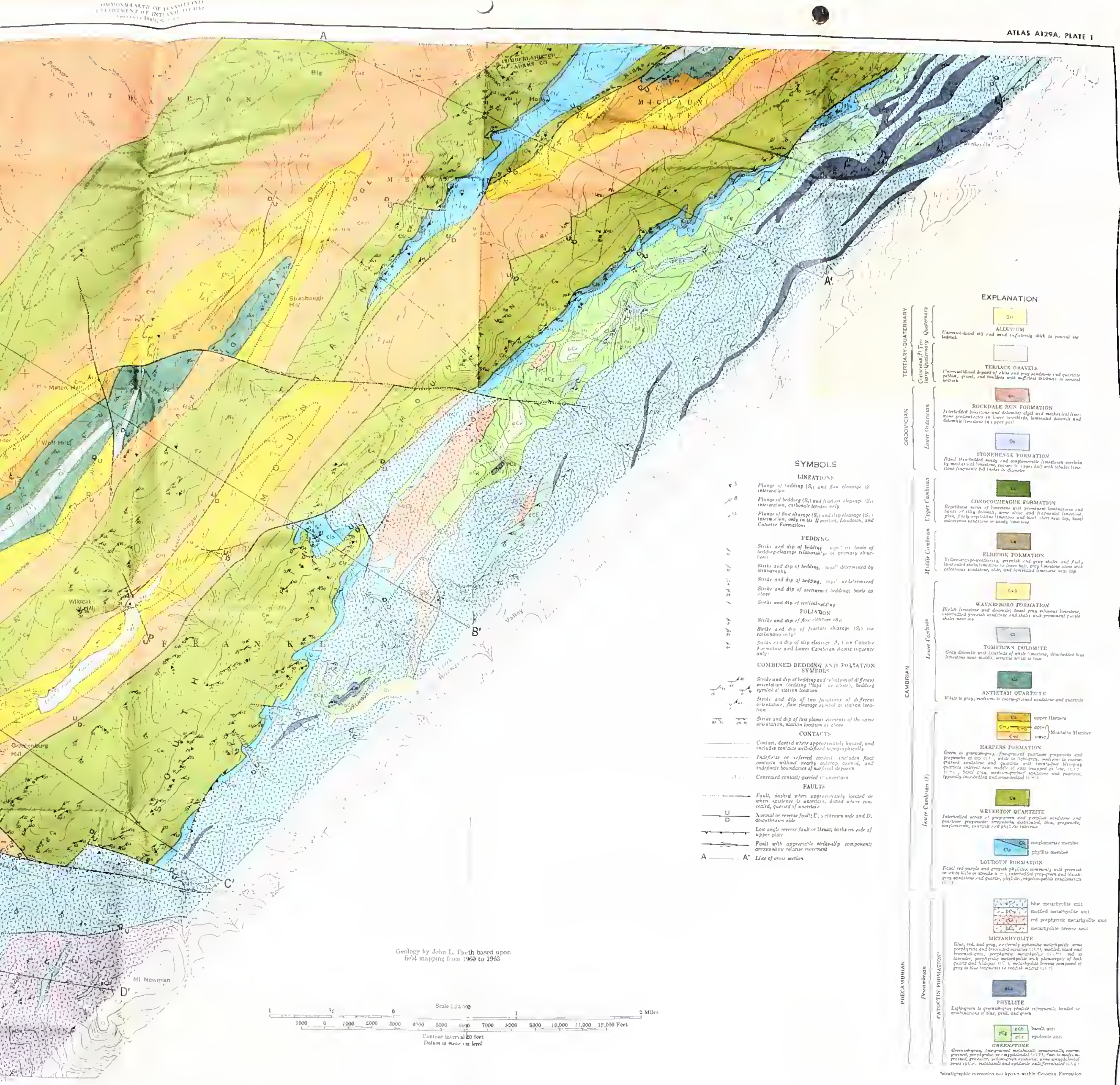

EXPLANATION

TERTIARY-QUATERNARY

Quaternary

Qal
ALLUVIUM
Unconsolidated silt and sand sufficiently thick to conceal the bedrock

Cretaceous(?) Tertiary-Quaternary

TERRACE GRAVELS
Unconsolidated deposit of white and gray sandstone and quartzite pebbles, gravel, and boulders with sufficient thickness to conceal bedrock

ORDOVICIAN

Lower Ordovician

Orr
ROCKDALE RUN FORMATION
Interbedded limestone and dolomite; algal and mechanical limestone predominates in lower two-thirds, laminated dolomite and dolomitic limestone in upper part

Os
STONEHENGE FORMATION
Basal thin-bedded sandy and conglomeratic limestones overlain by mechanical limestone, common in upper half with tabular limestone fragments 1-3 inches in diameter

CAMBRIAN

Upper Cambrian

Cc
CONOCOCHEAGUE FORMATION
Repetitious series of limestone with prominent laminations and bands of silty dolomite, some oolite and fragmental limestone, pink, finely crystalline limestone and black chert near top, basal calcareous sandstone or sandy limestone

Middle Cambrian

Ce
ELBROOK FORMATION
Yellow-orange-weathering, greenish and gray shales and finely laminated shaly limestone in lower half, gray limestone above with calcareous sandstone, oolite, and laminated limestone near top

Lower Cambrian

Cwb
WAYNESBORO FORMATION
Bluish limestone and dolomite; basal gray siliceous limestone, interbedded greenish sandstone and shales with prominent purple shales near top

Ct
TOMSTOWN DOLOMITE
Gray dolomite with interbeds of white limestone, thin-bedded blue limestone near middle, sericitic schist at base

Ca
ANTIETAM QUARTZITE
White to gray, medium- to coarse-grained sandstone and quartzite

Lower Cambrian (?)

Ch — upper Harpers
Cmu — upper
Cmb
Cml — lower
Montalto Member

HARPERS FORMATION
Green to greenish-gray, fine-grained quartzose graywacke and graywacke at top (Ch), white to light-gray, medium- to coarse-grained sandstone and quartzite with laminated blue-gray quartzite interval near middle of unit (mapped as line, Cmb) (Cmu), basal gray, medium-grained sandstone and quartzite, typically thin-bedded and cross-bedded (Cml)

Cw
WEVERTON QUARTZITE
Interbedded series of gray-green and purplish sandstone and quartzose graywacke; irregularly distributed, thin, graywacke, conglomerate, quartzite and phyllite intervals

Clc — conglomerate member
Clp — phyllite member

LOUDOUN FORMATION
Basal red-purple and grayish phyllite, commonly with greenish or white blebs or streaks (Clp), interbedded gray-green and bluish-gray sandstone and quartzite, phyllite, rhyolite-pebble conglomerate (Clc)

PRECAMBRIAN

Precambrian

CATOCTIN FORMATION*

pCbl — blue metarhyolite unit
pCm — mottled metarhyolite unit
pCr — red porphyritic metarhyolite unit
pCx — metarhyolite breccia unit

METARHYOLITE
Blue, red, and gray, extremely aphanitic metarhyolite; some porphyritic and brecciated varieties ([illegible]), mottled, black and brownish-gray, porphyritic metarhyolite ([illegible]), red to lavender, porphyritic metarhyolite with phenocrysts of both quartz and feldspar ([illegible]), metarhyolite breccia composed of gray to blue fragments in reddish matrix ([illegible])

pCs
PHYLLITE
Light-green to greenish-gray phyllite infrequently banded in combinations of blue, pink, and green

pCg
pCb — basalt unit
pCe — epidotite unit

GREENSTONE
Greenish-gray, fine-grained metabasalt; occasionally coarse-grained, porphyritic, or amygdaloidal (pCb), fine- to medium-grained, granular, yellow-green epidotite, some amygdaloidal zones (pCe), metabasalt and epidotite undifferentiated (pCg)

*Stratigraphic succession not known within Catoctin Formation

SYMBOLS

LINEATIONS

5 — *Plunge of bedding (S₀) and flow cleavage (S₁) intersection*

8 — *Plunge of bedding (S₀) and fracture cleavage (S₁) intersection, carbonate terrane only*

14 — *Plunge of flow cleavage (S₁) and slip cleavage (S₂) intersection, only in the Weverton, Loudoun, and Catoctin Formations*

BEDDING

Strike and dip of bedding, "tops" on basis of bedding-cleavage relationships or primary structures

Strike and dip of bedding, "tops" determined by stratigraphy

Strike and dip of bedding, "tops" undetermined

Strike and dip of overturned bedding; basis as above

Strike and dip of vertical bedding

FOLIATION

Strike and dip of flow cleavage (S₁)

Strike and dip of fracture cleavage (S₁) (in carbonates only)

Strike and dip of slip cleavage (S₂) (in Catoctin Formation and Lower Cambrian clastic sequence only)

COMBINED BEDDING AND FOLIATION SYMBOLS

Strike and dip of bedding and foliation of different orientation (bedding "tops" as above), bedding symbol at station location

Strike and dip of two foliations of different orientation, flow cleavage symbol at station location

Strike and dip of two planar elements of the same orientation, station location as above

CONTACTS

Contact, dashed where approximately located, and includes contacts well-defined topographically

Indefinite or inferred contact; includes float contacts without nearby outcrop control, and indefinite boundaries of surficial deposits

Concealed contact; queried if uncertain

FAULTS

Fault, dashed where approximately located or where existence is uncertain, dotted where concealed, queried if uncertain

U
D
Normal or reverse fault; U, upthrown side and D, downthrown side

Low angle reverse fault or thrust; barbs on side of upper plate

Fault with appreciable strike-slip component; arrows show relative movement

A — A' *Line of cross section*

Geology by John L. Fauth based upon
field mapping from 1960 to 1963

Scale 1:24 000
1 ½ 0 1 2 Miles
1000 0 1000 2000 3000 4000 5000 6000 7000 8000 9000 10,000 11,000 12,000 Feet
Contour interval 20 feet
Datum is mean sea level

GEOLOGIC MAP OF PARTS OF THE SCOTLAND, CALEDONIA PARK, AND ARENDTSVILLE QUADRANGLES, PENNSYLVANIA.